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A brief Abstract is required at the beginning of the paper. It should indicate the scope of the paper and give the principal results, and should be suitable for reproduction by abstracting journals as it stands.

All matter to be printed in italic type (e.g., generic and specific names) must be underlined. Headings are not to be underlined.

The "Style Book" of the N.Z. Government Printing Office may be used as a general guide to spelling, capitalisation, hyphenation, punctuation, abbreviations, etc.

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Tabular matter costs more to print than letterpress and should be kept to a minimum. Tables should be arranged to fit upright on the page, should be numbered consecutively in arabic numerals, and should have a brief explanatory title typed above each one. Units of measurement should be placed in parentheses at the head of the column, not in the body of the table. Descriptive notes should be kept to a minimum, and Government Style Book or SI abbreviations used wherever possible. Numerous small tables should be avoided. When submitted, all MSS. tables must be placed together in a section following the References. Mention in the text should be to the Table number, e.g., *see* Table 3.

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These are to be numbered consecutively in arabic numerals, regardless of whether they are half-tones (photographs, drawings) or line blocks (graphs, maps, etc.). The figure number should not be given on the figure itself, unless the figure is composite. Each must be referred to in the text, and the order of first reference must be in accordance with the consecutive numbers of the figures. Only such figures as are essential to elucidate the text can be published.

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# SUBDIVISION AND PETROLOGY OF THE MESOZOIC ROCKS OF COROMANDEL (MANAIA HILL GROUP)

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ABSTRACT

The original unconformable threefold subdivision of the sedimentary Mesozoic basement rocks of Coromandel is redefined as two conformable formations (Moehau and Tokatea Hill) within Manaia Hill Group on the basis of gradational detrital content and structural continuity; the new sequence is inverted with respect to previous subdivision. Standard sections are described for both formations (Moehau 520–610 m minimum; Tokatea Hill 530 m maximum).

The older formation (Moehau) is characterised by a lithic volcanic greywacke and sub-greywacke suite derived from calc-alkaline volcanic and plutonic rocks, and minor sedimentary rocks, whereas the younger formation (Tokatea Hill) is a feldspathic greywacke suite, almost devoid of volcanic detritus except near the base, derived from a more mature calc-alkaline plutonic landmass with minor sedimentary rocks.

Moehau Formation and the lower part of the Tokatea Hill Formation are slope-deposited turbidites, whereas the upper Tokatea Hill Formation resulted from stable deposition in deeper water.

Moehau Formation locally includes prehnite, analcime, actinolite, and zeolite vein minerals and has a chloritic matrix; Tokatea Hill rocks contain none of these minerals and typically have a sericitic matrix.

The structure of the Manaia Hill Group is relatively simple, the fold axes striking north-west to north-north-west with the Tokatea Hill Formation occupying the core of a syncline.

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## INTRODUCTION

Because "greywackes" form the oldest rocks exposed in the Coromandel goldfield (Fig. 1), early geological reports have tended to be more detailed in their description and subdivision of this rather monotonous and sparsely fossiliferous sequence, than of comparable Mesozoic or older geosynclinal sequences elsewhere in New Zealand.

Initial work was centred in the Tokatea–Tiki region to the north-east and south of Coromandel (Hector 1870; Hutton 1870; Cox 1881), where a belt of rock variously described as felsite (Cox 1882; Hector 1882), felsite tuffs (Hutton 1888; McKay 1897), yellow-grey pyritous mudstone (Park 1897), "ceratophyre" (Maclaren 1900), soda felsites (McKay 1904), or altered rhyolite and rhyolitic tuffs (Sollas and McKay 1905; Fraser and Adams 1907; Smale 1962) excited considerable interest. Opinion varied amongst the various workers as to whether these rocks intruded or were interbedded with the "slates" which made up the bulk of the basement rocks.

Further north near Colville, McKay (1886) and later workers described the rocks as slates, sandstones, greywackes, and "blue concretionary limestone" without conglomerates or volcanic detritus. However, workers in the Manaia–Waiau–Matawai region south of Coromandel found prominent sandstone and conglomerate beds, the pebbles and grains of which were predominantly of igneous origin (Maclaren 1900). Furthermore, this igneous detritus was said to be identifiable with the "felsites" or "altered rhyolites", etc. of Tokatea Hill, as well as andesite and diorite intrusive rocks of the Colville–Moehau region (McKay 1904), and andesite/dacite intrusive rocks from the Tokatea–Tiki–Pukewhau belt (Sollas and McKay 1905) which up to this time had been placed in the oldest Tertiary volcanic series overlying the basement.

Prior to 1907, various broad ages had been assigned to the basement rocks dependent upon the author's individual preference for correlation with fossiliferous rocks of the South Island of New Zealand. The slates and "felsites", etc. of Tokatea Hill were variously assigned to a Paleozoic (Cox 1881) or upper Devonian (Sollas and McKay 1905) Te Anau Series, or to a Devonian (Park 1897) or Carboniferous (McKay 1897, 1904) Maitai Series. The "slates" and associated rocks of Colville, along with the dioritic and andesitic intrusive rocks of Moehau, were assigned to the lower Carboniferous (McKay 1886); or to a Jurassic (Park 1893) or Devonian (Park 1897) or Carboniferous (McKay 1897; Sollas and McKay 1905) Maitai Series. The igneous conglomerates, sandstones, and siltstones of the Matawai–Waiau–Manaia block were correlated with the Wairoa Series of Triassic age (McKay 1904; Sollas and McKay 1905).

Thus there was considerable confusion as to what rocks should actually constitute the basement of Coromandel (*see* Table 1 *between* pp. 206–7). The net result of the early studies was the formulation of a three fold stratigraphy by McKay (*in* Sollas and McKay 1905), which was later amended by Fraser and Adams (1907) (Table 1) who attempted to trace boundaries and assigned more realistic age limits based on the discovery of Jurassic fossils at Manaia Hill from where poorly preserved ?*Monotis* had been previously identified by McKay.

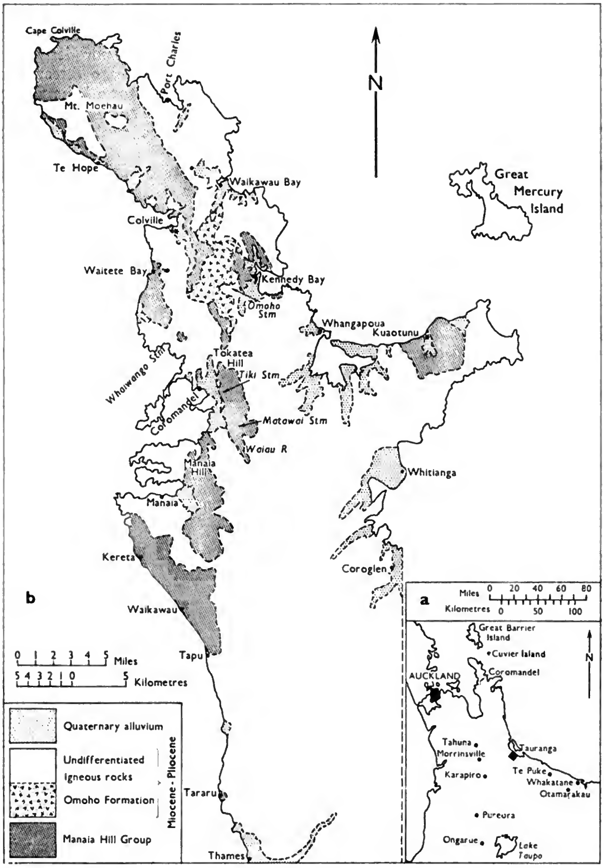



FIG. 1—
(a) General locality map; other Manaia Hill Group localities to south and south-east of Coromandel indicated by large dots.
(b) Distribution of Manaia Hill Group at Coromandel and the location of rhyolitic Omoho Formation, the narrow southern end of which was once included within the basement rocks.

Since this date, most workers have agreed that Moehau and Manaia Hill Series are identical in character and structurally continuous in outcrop (Smale 1962; Skinner 1962), but Smale (1962) insisted that these two series were distinct from a Tokatea Hill Series of acidic volcanic derivation although he examined only the limited outcrop in Whaiwango Stream. In recent years the term "Manaia Hill Group" has been used for all the poorly fossiliferous eastern lithic-volcanic facies of the greywacke suite in the south Auckland–Coromandel region. Originally the term was applied only to rocks of the old Manaia Hill and Moehau Series (Kear 1955), but has since been extended to include the old Tokatea Hill Series (Schofield 1967). Within this group, "Moehau Formation" has been applied to the basement rocks of Great Barrier Island, Cuvier Island and Cape Colville–Moehau (Thompson 1960).

The present paper will use the terminology of Schofield (1967) but will introduce "Tokatea Hill Formation" for much of those rocks once classified as Tokatea Hill Series, and will extend "Moehau Formation" to cover the old Moehau and Manaia Hill Series (Fig. 2). Thus the Manaia Hill Group in the Coromandel region, as now defined, includes the Moehau and Tokatea Hill Formations. Sample numbers are assigned by Department of Geology, University of Auckland; grid references refer to Fig. 2 and 5, and to NZMS 1 sheets N35–36, 39, 40, 43, and 44.

## MOEHAU FORMATION

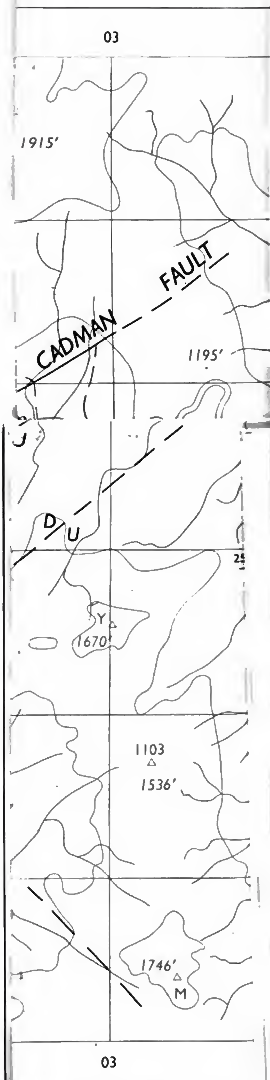
### *Definition and Thickness*

Moehau Formation includes all those rocks previously mapped as Manaia Hill and Moehau Series, with the addition, because of their structure and petrography, of part of those rocks mapped as Tokatea Hill Series. The classical type locality has been an old quarry on Manaia Hill at N43/979635 where a conglomerate with upper Jurassic fossils (Kear 1955) is exposed. However, it is proposed that the section in Matawai Stream between N44/027668 and N44/047675 (Fig. 2, 3 and 4) be defined as a standard section, for the faulted anticlinal sequence presents the most continuous exposure of the formation, and includes the contact with overlying Tokatea Hill Formation. Although some 520 m to 610 m of strata are exposed, the possible total thickness is perhaps considerably greater as other exposures show evidence for repetition, overturning, faulting, and slumping.

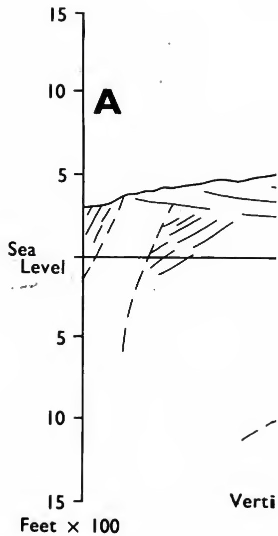
#### *Standard Section*

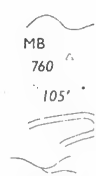
The standard section consists of an isoclinal, asymmetric anticline with its axial plane dipping at 65° to the west. It is downfaulted along the Matawai Fault so that the eastern limb is repeated and includes the overlying Tokatea Hill Formation. The western limb, which passes into at least two small anticlines with almost vertical axial planes, is separated from Miocene andesites (Beesons Island Volcanics) by the Coromandel Fault (Fig. 3). The succession upstream (eastwards) is:

ks




ia. For geological legend *see* Fig. 3.

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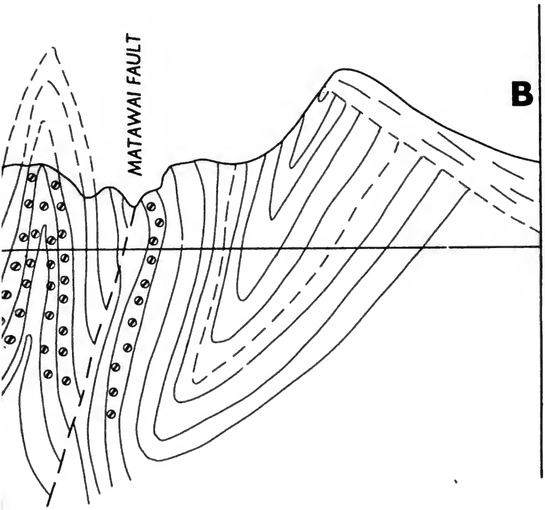


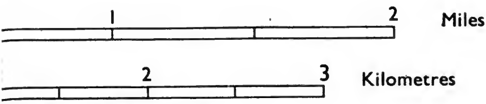


## ICAL SYMBOLS

cal

Faults — certain

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20

U

D

probable

obscured

urned

cal

Dip on fault plane, direction of dip slip and strike slip and plunge on slickensides indicated.

Table 2—Modal Composition

(See opposite page for l

| No. | 1a | 1b | 2a | 2b | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 1 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Grade | css–mcg | css–mcg | css–mcg | css–mcg | css | m-c–ss | m-c–ss | zs | m-c–ss | css | css | : |
| No. of Points | 2000 | 2000 | 2000 | 1000 | 1000 | 1000 | 1000 | 500 | 1000 | 1000 | 1500 | 5 |
| Q | 7.1 | 6.8 | 4.4 | 5.0 | 5.7 | 8.9 | 10.9 | 25.0 | 7.9 | 10.8 | 6.4 | 3: |
| P | 5.9 | 12.6 | 13.8 | 29.1 | 14.1 | 15.1 | 15.1 | 4.2 | 22.1 | 11.8 | 15.6 | ( |
| O | 1.1 | 1.6 | 0.1 | 0 | 0 | 2.3 | 3.8 | 5.8 | 0 | 2.6 | 2.1 | ( |
| rfv | 52.0 | 56.2 | 61.8 | 54.5 | 63.7 | 54.7 | 57.7 | 28.0 | 40.0 | 48.5 | 51.0 | ( |
| rf | 13.6 | 5.2 | 6.8 | 2.2 | 6.8 | 3.2 | 1.2 | 1.6 | 3.2 | 9.2 | 7.8 | 1 |
| Acc | 0.2 | 0 | 0.1 | 0 | 0 | 0.6 | 0 | 0.2 | 1.5 | 0.6 | 0.9 | ( |
| Mat | 20.1 | 17.6 | 13.0 | 9.2 | 9.7 | 15.2 | 11.3 | 35.2 | 25.3 | 16.5 | 16.2 | 5: |
| | Moehau Formation WEST OF SYNCLINE | | | | | | | | | | | |
| Acc | b | b | h | h | b/m<br>h | b/a<br>A/c<br>m | e/a<br>c/p<br>A | m | e/F | c/p<br>b/z<br>F | p/e | b<br>1 |
| Rock type | g<br>L | g<br>L | sg<br>L | sg<br>L | sg<br>L | ‡g/sg<br>L | sg<br>L | g<br>L | g<br>L | g/sg<br>L | g/sg<br>L | : |

*felsic trend; predominantly lithic †equally lithic and felsic ‡greywacke-subgreywacke

of Manaia Hill Group

ey to symbols)

| ɔ | 11 | 12 | 13 | 14 | 15 | 16 | 17 | 18 | 19 | 20 | 21 | 22 | 23 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| ≥ | zs | zs fss | zs \| fss | zs | zs \| ss | zs \| mss | zs \| fss | zs \| fss | zs \| mss | zs \| ss | css mcg | zs \| fss | zs mss |
| O | 500 | 500 | 500 | 500 | 1000 | 1000 | 1000 | 1000 | 1000 | 500 | 2000 | 500 | 1000 |
| .0 | 20.6 | 31.6 | 34.4 | 28.1 | 14.1 | 25.1 | 11.8 | 18.2 | 9.5 | 18.0 | 2.5 | 16.2 | 3.3 |
| .6 | 1.2 | 3.0 | 21.8 | 18.6 | 21.2 | 14.4 | 26.3 | 18.8 | 20.3 | 22.6 | 10.7 | 15.8 | 14.8 |
| | 0 | 2.4 | 0.6 | 2.8 | 0 | 2.5 | 1.8 | 2.9 | 4.9 | 0.4 | 0.5 | 4.8 | 0 |
| .6 | 0 | 1.6 | 3.0 | 3.6 | 9.6 | 17.1 | 29.0 | 14.2 | 20.7 | 14.6 | 63.5 | 23.4 | 34.4 |
| .O | 0.6 | 1.8 | 1.8 | 2.0 | 3.0 | 4.2 | 6.5 | 2.9 | 3.9 | 5.2 | 14.3 | 1.0 | 3.0 |
| .8 | 1.2 | 2.4 | 0.8 | 1.2 | 0.4 | 0.8 | 0.5 | 0.2 | 1.1 | 1.2 | 0 | 0 | 0.2 |
| .0 | 76.4 | 57.2 | 37.6 | 43.7 | 51.7 | 35.9 | 24.1 | 42.8 | 39.6 | 38.0 | 8.5 | 38.8 | 44.3 |
| | Tokatea Hill Formation SYNCLINE | | | | | Moehau Formation EAST OF SYNCLINE | | | | Moehau Formation scattered localities | | | |
| m I | b/m t/z | b/m z/A g | b/m/g z/z/A S? | m/h b/e A/p z/H | c | m/b e/z c/P | b m | m A z | m z e | b m/a z/g e/h | b | e/z | P/z |
| : | g f | g f | g f | g f | g f | g L *(f) | g L (f) | g f §(L) | g †L/f | g f (L) | sg L | g L (f) | g L |

boundary composition §lithic trend; predominantly felsic

## KEY TO TABLES 2, 3, AND 4

g = greywacke
sg = subgreywacke
L = lithic
f = feldspathic
} rock type after Pettijohn (1957)

Accessory types

| | | | |
|---|---|---|---|
| a | actinolite | A | apatite |
| b | biotite | | |
| c | chlorite | P | pennine |
| e | epidote | | |
| g | garnet | | |
| h | hornblende | H | hypersthene |
| m | muscovite | | |
| p | pyroxene | F | pyrite |
| t | tourmaline | S | sphene |
| z | zircon | | |

Grain types

| | |
|---|---|
| Q | quartz |
| P | plagioclase |
| O | potash feldspar |
| rfv | volcanic rock fragments |
| rf | non-volcanic rock fragments |
| Acc. | accessories |
| Mat. | matrix |

| Grade Range—Average | | mm | Symbol |
|---|---|---|---|
| matrix | less than | 0.01 | |
| siltstone | | 0.03 – 0.075 | zs |
| sandstone | fine | 0.075– 0.15 | fss |
| | medium | 0.15 – 0.3 | mss |
| | coarse | 0.3 – 1.0 | css |
| conglomerate | fine | 1.0 – 2.0 | mcg |
| | greater than | 2 | cg |

| Mode No. | Auckland University Sample No. | NZMS 1 Grid Reference |
|---|---|---|
| 1a/b | 11670 | 44/023685 |
| 2a/b | 11673 | ,, |
| 3 | 11674 | ,, |
| 4 | 11675 | 44/042673 |
| 5 | 11677 | 40/020702 |
| 6 | 11686 | 44/047674 |
| 7 | 11695 | 40/014721 |
| 8 | 11716 | 44/034670 |
| 9 | 11717 | 44/036670 |
| 10 | 11728 | 40/017770 |
| 11 | 11730 | 44/033692 |
| 12 | 11739 | 40/024715 |
| 13 | 11740 | 40/026715 |
| 14 | 11733 | 44/033685 |
| 15 | 11731 | 40/029712 |
| 16 | 11679 | 44/033685 |
| 17 | 11689 | 40/028716 |
| 18 | 11690 | 40/029716 |
| 19 | 11693 | 40/030716 |
| 20 | 11699 | 43/979634 |
| 21 | 11700 | 43/979635 |
| 22 | 11708 | 40/066805 |
| 23 | 11714 | 35/834057 |

Samples 1a/b and 2a/b represent two counts per sample:
a = counting rock fragments as whole fragments
b = counting porphyritic minerals of rock fragments as mineral fragments.

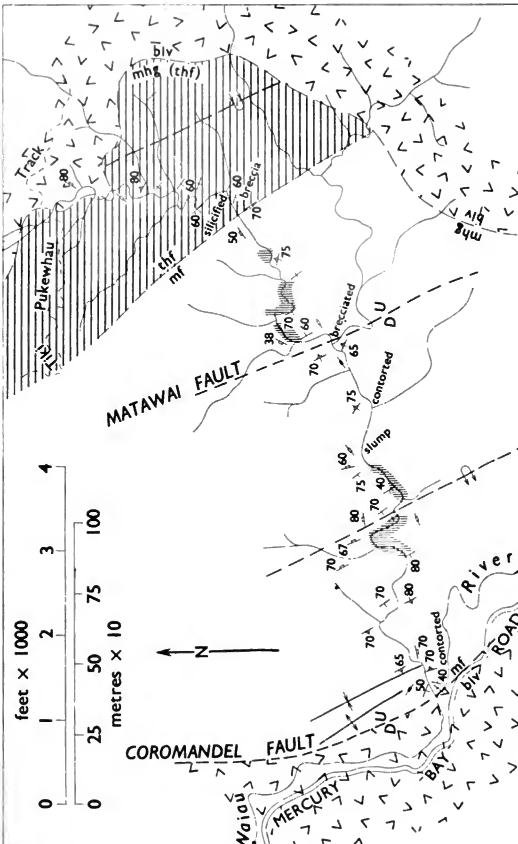


FIG. 4—Plan of standard Moehau section at Matawai Stream. Symbols as for Fig. 2; hachuring marks sandstone-conglomerate outcrops; biv = Beesons Island Volcanics; mhg = Manaia Hill Group; thf = Tokatea Hill Formation; mf = Moehau Formation.

Beesons Island Volcanics

| | | | | |
|---|---|---|---|---|
| — — — | — — — Coromandel Fault — — — | | | |
| N44/02706680 | irregular folded siltstone and argillite | | | |
| 02906690 | well bedded siltstone and argillite | 312 m thick | | West Limb |
| 03306695 | sandstone and conglomerate | 195 m | | West Limb |
| 03306690 | well bedded siltstone and argillite | 9 m | | West Limb |
| — — — | — — — axis of fold — — — | | | |
| | well bedded siltstone and argillite | 9 m | | East Limb |
| 03656695 | sandstone and conglomerate | 189 m | | East Limb |
| 03806705 | well bedded siltstone and argillite | 236 m | | East Limb |
| 04106715 | irregular siltstone and argillite | 49 m | | East Limb |
| — — — | — — — Matawai Fault — — — | | | |
| 04206720 | irregular siltstone and argillite | 91 m | | East Limb |
| 04256730 | sandstone and conglomerate | 199 m | | East Limb |
| 04506740 | argillite, siltstone, minor sandstone | 220 m | | East Limb |
| 04706745 | — — — — — — — — | | | |

Tokatea Hill Formation

## *Content*

The rocks of Moehau Formation are typically hard, well indurated, dark blue-grey conglomerate to sub-fissile argillite. No interbedded volcanics are known, chert is rare (Skinner 1962), and although limestones have been reported (McKay 1886) their presence has not been confirmed.

The examination of 70 thin sections and the modal analyses of seventeen representative samples have indicated that, following Pettijohn (1957), the Moehau Formation is typified by lithic volcanic greywackes and sub-greywackes (*see* Table 2 *facing* p. 207).

### *Conglomerate*

Both "argillite-flake breccia" (chipwacke) and normal roundstone conglomerate are present, the former consisting of angular, lensoid fragments of argillite or fine siltstone (50 cm × 5–7 cm) embedded in coarse siltstone, sandstone, or grit, and the latter consisting of rounded to sub-rounded volcanic pebbles (up to 30 mm in diameter) with very subordinate sedimentary or metamorphic pebbles, and very little matrix, the interstices being filled by smaller pebbles, sub-angular to sub-rounded quartz, plagioclase and orthoclase, and accessories such as biotite, epidote and muscovite. The density averages 2·70 to 2·73.

Both types of conglomerate occur together (Manaia Hill Quarry), but chipwacke is also associated with sandstones (Tupa Stream) and intraformational slump deposits (Tiki Stream at Quarry Road), and the volcanic-derived conglomerate grades into and contains lenses and beds of coarse sandstone. Large areas in which bedding and hence thickness are not obvious occur at Manaia Hill, Mill Stream, and Tupa Stream. Much of the sandstone and siltstone of Tupa Stream west of the conglomerate, and south of the Manaia Hill quarry show gradational affinities with Tokatea Hill Formation. Likewise along the Tiki–Matawai belt the conglomerate member occurs some 266 m below the Tokatea Hill Formation and the intervening sandstones and siltstones show a gradational change in detrital content towards the boundary.

*Sandstone*

Sandstone is gradational with conglomerate, the main difference being a higher percentage of crystalline detritus and matrix in the former due to breakdown of larger rock fragments. Density measurements ranged from 2·54 to 2·66. Bedding is emphasised by lenses or layers of siltstone, but generally the sandstones are massive and well jointed, with an unsorted porphyroclastic texture of large sub-rounded to sub-angular crystalline, and rounded to sub-rounded lithic detritus embedded in a dark, chloritic, argillaceous matrix containing fine angular fragments of quartz and plagioclase which grade into the matrix itself.

*Siltstone and Argillite*

Interbedded siltstone and argillite forms the bulk of the Moehau Formation. Beds seldom reach 25 cm thick and average 7·5 cm. Siltstone densities range from 2·68 to 2·73, finely bedded siltstone and argillite gave density values of 2·54 and 2·57, but one argillite sample had a density of 2·75.

Texturally and mineralogically siltstone resembles sandstone but has an even higher proportion of matrix and quartz. Argillite has a fine texture of angular quartz fragments aligned with sericite, chlorite and clay in skeins parallel to the bedding.

## *Petrography*

*Quartz*

4% to 25% (average 10%) of the detritus. Generally sub-angular but rounded to sub-rounded in coarser rocks. Sharply angular fragments result from shattering (AU11677, AU11695); most grains are crushed and all are strained. Pressure solution suturing is rare (AU11683) as grains are seldom in contact, but large secondary overgrowths and interconnecting plumose quartz sprays are common (AU11670). Shatter cracks are filled by fine quartzo-feldspathic matrix and calcite, the latter replacing the quartz (AU11673). Inclusions of muscovite flakes (AU11670), apatite (AU11675, AU11695) and epidote (AU11708, AU11717) usually present.

*Plagioclase*

6% to 30% (average 16%) of detritus. Generally fairly constant composition from $An_{28-44}$, but rare more basic plagioclase ($An_{50}$–AU11699) also found in volcanic fragments. Sub-angular to sub-rounded and irregularly twinned. Characteristically, both perfectly fresh and strongly sericitised (AU11670, AU11675) or albitised (AU11677, AU11699, AU11700) plagioclase common to most samples. Range haphazardly in size from more than 5 mm (conglomerates) down to 0.01 mm (argillites). Generally includes apatite prisms (AU11699).

*Orthoclase*

0% to 6% (average 2%) of detritus. Confirmed by sodium cobaltrinitrite/HF staining (after Chayes 1952). Very irregular and erratic occurrences; adjacent samples differ considerably (Table 2: AU11670, AU11673), as it may be concentrated in patches (AU11708) or certain lamellae (AU11713). Sub-rounded, commonly sericitised micro perthitic grains, rarely simply twinned (AU11675), and seldom larger than 0·15 mm. Perthite not uncommon (AU11670, AU11677), but antiperthite (AU11675) and microcline (AU7552) are rare; quartz (AU11675) or zircon (AU11693) are rarely included.

*Volcanic Rock Fragments*

15% to 65% (average 41%) of detritus. Rounded to sub-rounded grains of all sizes between 30 mm and matrix (0·05–0·01 mm). Diffuse, indistinct grain boundaries. Plagioclase microlites $An_{28-44}$, and phenocrysts $An_{52}$ (AU11700); usually severely altered. Ferromagnesians represented by occasional resorption pseudomorphs of hornblende (AU11717), or chlorite pseudomorphs after pyroxene (AU11700) or biotite (AU11698). Both extremely altered (epidote-pyrite-chlorite–AU7591; chlorite–AU11699) and relatively fresh fragments co-existing in most samples.

Trachytic and often porphyritic andesite-dacite fragments predominate, but coarsely porphyritic grains more usual in the conglomerates. Quartz phenocrysts are rare (AU11670). Other less common rock types include rhyolites with devitrified, potassic glass (AU11670), obsidian (AU11715), perlitic and spherulitic rhyolites (AU11670, AU11700, AU11717), ignimbrite (AU7457), quartz-bearing microlitic and porphyritic feldspathic rhyolites (AU11670, AU11675), and fine grained feldspathic rhyolites (AU11700, AU11716). Dark, finely microlitic, opaque-rich basalt is a rare constituent (AU11699).

*Other Rock Fragments*

1% to 14% (average 5%) of detritus. Most fragments sub-rounded to sub-angular argillite and siltstone with minor rounded chert (AU11670) and well rounded greywacke. Fragments of probable metamorphic origin represented by occasional well rounded pebbles of tightly sutured quartzite, and sub-rounded, hornfels-like fragments with assemblages such as quartz-muscovite-relict detrital plagioclase (AU11670), biotite-quartz (AU11674), and muscovite-epidote-quartz (AU11706). Pebbles of plutonic rock are rare (quartz-orthoclase and quartz-plagioclase–AU11702) to common (quartz-orthoclase myrmekite–AU11674, and micrographic intergrowth–AU11708, AU7552); serpentine is very rare (AU11675). Fragments of vein quartz (AU11670), calcite rhombs (AU11673) and quartz-epidote or epidote-plagioclase (AU11717) are locally abundant.

*Accessories* (Table 3)

0% to 1·5% of the detritus, most of which is biotite and muscovite. Biotite, very dark brown (AU11670, AU11675, AU11712, AU11715) or golden brown (AU11687, AU11696, AU11698), occurs as both small and very large (AU11670) bent and twisted flakes (AU11699). Less common yellowish, ploechroic muscovite (AU11701–2) occurs as many small flakes (AU11698), large plates (AU11675), or sprays (AU11681).

Epidote is both granular and euhedral; it accounts for most of the 1·5% accessories of AU11695.

Greenish-brown hornblende (AU11698–9) is occasionally present as small, angular, chloritised fragments (AU11677) or rarely as large corroded and leached subhedra (AU11675). Green actinolitic hornblende is much more abundant (AU11675, AU11699).

Augite occurs as small chloritised and eroded fragments (AU11677, AU11702–3), or large (0·75 mm) fresh angular subhedra with chloritised margins and calcified joints (AU11716–7).

The accessories also include brownish-green chlorite, pennine, angular zircons (0·03–0·05 mm; rarely 0·1 mm), apatite, colourless to yellow rounded garnet, eroded and rounded "cubes" of pyrite and, at the top of Moehau Formation sequence, sphene.

TABLE 3—Accessory Minerals of Moehau Formation

Symbols at head of columns are explained in Key to Table 2, *see* p. 207 (r = rare. x = abundant; v = very abundant)

| | a | b | c | e | g | h | m | p | t | z | A | P | H | F | S |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 11717 | | | | x | | | | | | | | x | | | |
| 11716 | | r | x | | | | | | | x | | x | | r | |
| 11715 | | x | | x | | | | | | | | | | | |
| 11714 | | | | | | | | | | | r | r | | | |
| 11713 | | | | r | | | | | | | | | | | |
| 11712 | | x | | | | | | | | | | | | | |
| 11711 | | | | | | | x | | | | | | | | |
| 11708 | | | | x | | | | | | r | | | | | |
| 11705 | | | | r | | | x | | | r | | | | | |
| 11703 | | r | | | | | | r | | | | | | | |
| 11702 | | | | v | | | x | r | | | | | | | |
| 11701 | | x | | x | | | x | | | | | | | | |
| 11700 | | x | | | | | | | | | | | | | |
| 11699 | r | v | | r | x | x | x | | | x | r | | | | |
| 11698 | | v | | | r | r | v | | | | r | | | | |
| 11697 | | | | | | | | | | x | | | | | |
| 11696 | | x | | | | | x | | | | | | | | |
| 11695 | | | | v | | | | | | | | | | r | |
| 11694 | | v | | | | | v | | | | | | | | |
| 11693 | | | | r | | | x | | | r | | | | | |
| 11692 | | | r | | | | x | | | | | | | | |
| 11691 | | | | x | | | | r | | | | | | | x |
| 11690 | | | | | | | r | | | r | r | | | | |
| 11689 | | r | | | | | r | | | | | | | | |
| 11688 | | x | | | | | x | | | | | | | | |
| 11687 | | x | | | | | x | | | | | | | | |
| 11686 | | | | | | | x | | | | | | | | |
| 11685 | | | | | | | v | | | | | | | | |
| 11684 | | x | | | | | x | | | | | | | | |
| 11682 | | | | | | | v | | | | | | | | |
| 11681 | | | | | | | x | | | | | | | | |
| 11679 | | x | x | x | | | x | | | r | | r | | | |
| 11677 | r | | x | r | | x | | r | | | | | | | |
| 11676 | | x | | | | | x | | | | | | | | |
| 11675 | r | r | r | | | r | x | | | | | | | | |
| 11674 | | r | | | | r | r | | | | | | | | |
| 11673 | | | | | | r | | | | | | | | | |
| 11671 | | | | x | | | | | | | | | | | |
| 11670 | | x | | | | | | | | | | | | | |

*Matrix*

10% to 45% (average 24%) of bulk of rock. Typically consists of partly recrystallised quartz-feldspar with dark, isotropic, perhaps carbonaceous but certainly pyritic clay skeins containing quartz and feldspar fragments <0·01 mm (AU11670). Brown-green chlorite (AU11674), calcite (AU11673), and rarely opal (AU11695) are interstitial; "clinochlore" is commonly present as fine fibres within the matrix (AU11677, AU11699, AU11700, AU11708). The matrix of rocks with high orthoclase content stains with sodium cobaltinitrite in well defined patches (AU11675) or completely (AU11686). Sericite is rare except near the boundary with Tokatea Hill Formation (AU11675, AU11684, AU11688).

*Secondary Mineralogy*

Quartz, calcite, prehnite (AU11717–8), and laumontite-leonhardite (AU11713, AU11721–3) can be seen to replace plagioclase and matrix in localities near shear zones or fold axes. Analcime (AU11721–2), clinozoisite (AU11709–10), white chlorite (AU11711–2), epidote (AU11671, AU11726), chlorite and sericite (AU11727) replace plagioclase and matrix in or adjacent to quartz-veined shear zones.

## *Origin*

*Provenance*

It can be readily appreciated that the source landmass consisted largely of volcanic rocks, principally andesites and dacites, but also including rhyolites and minor basalts. That these were felspathic could be inferred from the relative absence of ferromagnesians; however, the corrosion of the fragments that are present, and the difference in abundance between included and detrital apatite, suggest that interstratal leaching could have occurred. The scarcity of ferromagnesians could, however, be due to the hydrothermal alteration that had obviously occurred prior to erosion of the source rocks. The abundance of epidote, the albitised and sericitised plagioclase associated with fresh plagioclase, the pyritic matrix, and the presence of detrital chlorite, pyrite, vein quartz, and rhombic calcite indicate that the alteration of the source rocks was widespread.

It is inferred from the amount of light-brown biotite and muscovite, the perthitic orthoclase, and the micrographic and myrmekitic intergrowths, that granodiorites and diorites were exposed to limited erosion. A coarse-grained texture is suggested by the absence of whole pebbles, although these have been reported from Great Barrier (Bartrum 1921) and Whakatane (Paltridge 1953). Some of the unaltered, coarse plagioclase, the micro-perthitic orthoclase, the quartz with apatite inclusions, and zircons could have a similar origin.

The occasional hornfels-like pebble, the rare quartzite, the detrital garnet, and particularly the dark brown, nearly black, biotite indicates the presence of low-grade metamorphosed "greywacke" sedimentary rocks in the terrain. Unmetamorphosed equivalents may be represented by siltstone, argillite, rare greywacke sandstone, and the chert pebbles, although these could also have a cannibalistic origin during sedimentation and be related to the intraformational source of the chipwackes. The rarity of chert in the exposed Moehau Formation but the abundance of chert in the detritus suggests that much of the sedimentary detritus was exogenous.

*Sedimentation*

In general, the beds have few sedimentary features except for lamination of siltstones, and monotonously alternating sequences of argillite and siltstone with occasional massive sandstone-conglomerate beds. Small load casts, seldom more than a few millimetres in amplitude, afford the main criterion for the determination of younging but graded bedding is locally conspicuous in the Tiki and Matawai streams. Intraformational slumping has obviously been widespread but is seldom of large amplitude. It is particularly common in the vicinity of, and below, sandstone and conglomerate beds where it is associated with penecontemporaneous flake-breccias, convolute bedding, and rafting of the overlying sediments.

*Environment of Deposition*

Moehau Formation resembles part of the Waipapa Formation of North Auckland (Hopgood 1956) and the Urewera Greywacke of Bay of Plenty (Healey, *et al.* 1964) which, however, were interpreted by Hopgood (1956) and Paltridge (1953) as products of deep-water geosynclinal sedimentation. There is little resemblance to the Alpine Facies of Wellman (1952) which is generally considered to be axial turbidites derived from granitic and high grade metamorphic rocks (Reed 1957), but Moehau Formation is comparable to Wellman's Hokonui Facies, a geosynclinal shelf facies characterised by lithic volcanic detritus.

The porphyroclastic texture with detrital fragments of all sizes down to matrix, gives to the rocks the appearance of having been rapidly and haphazardly "poured in" as a mud slurry. The angularity and freshness of even the hydrothermally altered detritus also implies rapid deposition. Poor sorting, loadcasts, convolute bedding, few fossils, slump structures, and rafting of competent sediments are characteristic features of rapid deposition by turbidity currents (Kuenen and Migliorini 1951). However, Kuenen (1953) also noted that "near source" deposits of the continental slope would be dominated by slumping and alternating non-graded sediments, whereas grading, and specially current bedding, would be very subordinate, these being characteristic of axial turbidites. Flake-breccias and even conglomerates can, together, be characteristic of turbidite redeposition (Kuenen and Saunders 1956).

Thus Moehau Formation has characteristics of turbidites in a "near source" (continental slope) environment, and has petrographic affinities with New Zealand shelf, rather than deep water axial, geosynclinal facies.

## *Structure*

*Folds and Bedding*

Throughout the length of the main divide exposures between Colville and Manaia, the strike of bedding planes maintains a fairly constant NW to NNW trend, interrupted only by drag effects on various faults. Dips tend to be greater than 65° to the west and many of the beds are overturned. On the basis of seven bed facings in Matawai Stream, three in Pukewhau Stream, and four in Cadman Stream, with, in addition, the continuity of a conglomerate-sandstone marker bed, the structure of Moehau Formation in this region has been determined as a faulted, assymetric anticline and syncline, with the

common limb overturned (Fig. 2, 3, 4,). Because of the steep dips, the folds must be iscoclinal and compressed, the axial planes striking approximately NNW and dipping steeply to the west; there is no distinct axial plunge. On Manaia Hill, the large area of conglomerate has rather variable W to NW strikes with steep northerly dips. The whole mass may be allochthonous, as the surrounding rocks of Tupa Stream and Awakanae River have relatively regular NW strikes except where interrupted by faults.

*Joints*

The massive sandstones and conglomerates have very clear joint patterns that are almost invariably symmetrically conjugate to the axial plane or bedding trend. Longitudinal, cross, and diagonal joints are particularly developed in Tupa and Matawai streams, where they correspond with a NW to NNW trend of fold axes with little plunge (Fig. 2, 3).

## TOKATEA HILL FORMATION

### *Definition and Thickness*

The Tokatea Hill Series of Sollas and McKay (1905) and Fraser and Adams (1907) was based on the inclusion of flow rhyolites and rhyolite derived sediments within the basement rocks. However, it has been conclusively proved by mapping along the strike that the flow rhyolites and rhyolitic tuffs are actually part of the Miocene volcanic sequence and belong to Omoho Formation (Schofield 1967; otherwise equivalent to Kapanga Rhyolite of Thompson 1966, or First Period Rhyolite of Fraser and Adams 1907). The sequence exposed by the road cuttings east of Tokatea Hill to Kennedy Bay (Fig. 5, *see colour insert between* pp. 206–7) shows the unconformable contact between the two formations, and the identity of the flow rhyolite and the rhyolite porphyry (or felsite) dike.

Apart from the rhyolites, there still remains a series of rocks corresponding to the "felsite tuffs" of Hutton (1888) and the "acid tuffs, tuffaceous mudstones, spotted felsitic grauwackes [sic] and quartz-sericite rocks" which previous authors were inclined to believe were of rhyolitic origin or even altered rhyolites. Instead, as Hutton (1888) noted, these are clastic rocks which are no more rhyolitic than Moehau Formation, but which are distinctly more felspathic.

A type locality for the original Tokatea Hill Series was not defined but typical exposures were reported for each rock type. The Tokatea and Royal Oak mines (N40/013766; 017772) exposed sections through the formation in which greywackes and slates were said to occur above and below the "felsites" (Cox 1882) or "felsite tuffs and ceratophyres" (Maclaren 1900) the whole dipping steeply to the west. The horizontal succession from west to east along the still accessible No. 7 level (the Tokatea Tunnell) of the Royal Oak mine was as follows:

| | |
|---|---|
| Tertiary volcanic rocks | 137 m |
| sandstones and shales | 30 m |
| felsite | 366 m |
| sandstones and shales | 76 m |

Similar successions were reported from the Cadman–Tiki–Pukewhau area (McKay 1904). Although the rhyolite flows and tuffs now form part of the Tertiary, there is still sufficient documentation to show that the remainder of the Tokatea Hill series includes strata which are unique to it and thus the name should be retained.

### *Standard Section*

It is proposed that the strata exposed in Petote Stream and its tributaries between N40/01957175 (just downstream of the junction with Aitken Stream) and N40/02757165 (approximately 90 m up the first large tributary from the east: Hut Creek) be defined as a standard section for Tokatea Hill Formation (Fig. 6). The complete eastern limb of a syncline is exposed, as well as the boundary with underlying Moehau Formation; as such the thickness (530 m) represents the maximum known for the formation. The succession eastwards (i.e., towards the base of the formation) is as follows:

| | | |
|---|---|---|
| | — — — axis of syncline — — — | |
| N40/01957175 | Interbedded thick or massive, light grey, fine sandstone, siltstone and mudstone | 119 m |
| 02207155 | Finely lamellate black pyritic argillite and grey siltstone | 38 m |
| 02257150 | Slumped, irregular, and sheared bedded grey siltstone and dark argillite | 78 m |
| 02407150 | Well bedded brown to black argillite, and grey siltstone and fine sandstone | 117 m |
| 02507150 | Narrow, irregular bedded, slumped zone | 3 m |
| | Thick bedded or massive, light grey, siltstone and mudstone with occasional thin, dark argillite lamellae | 33 m |
| 02557150 | Slumped, irregular, but well bedded grey argillite and siltstone with some dark grey argillite | 56 m |
| 02657155 | Well bedded light grey mudstone, siltstone and fine sandstone | 86 m |
| 02757165 | — — — — — — — — — | |
| | Moehau Formation | |

## *Lithologies*

Tokatea Hill strata are typified by light grey, medium hard to soft mudstones and siltstones that tend to be either massive or in beds up to 10 m thick. Densities range from 2·69 to 2·73. The lithologies weather easily, but when fresh (as in the Royal Oak mine) tend to be greenish; these are certainly the felsites and felsitic tuffs of previous workers. The light grey to green colouration is due to a high proportion of sericite and varying but small quantities of chlorite in the matrix.

The spotted mudstones and siltstones (spotted felsitic rocks of earlier workers) have a finer texture and a sericitic matrix, with almost spherical concentrations up to 2 mm in diameter of coarser plagioclase-quartz detritus (0·075 mm) in a chlorite-sericite matrix.

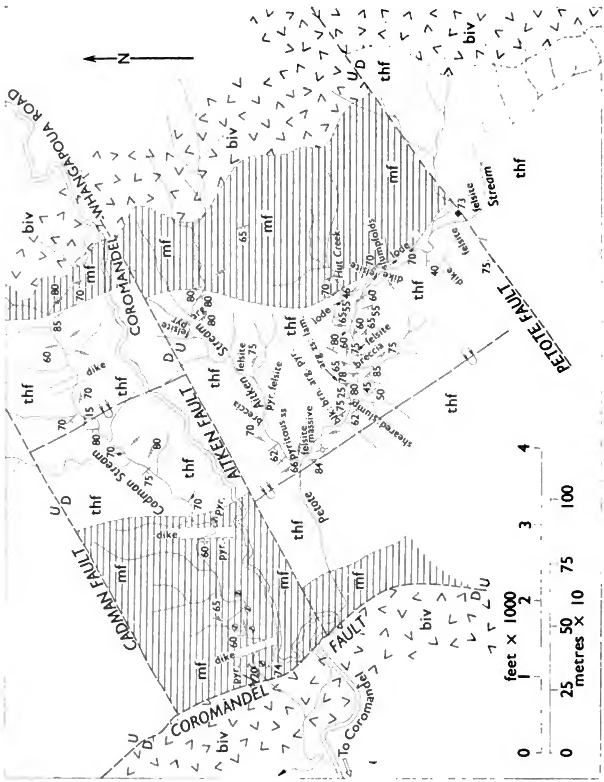


FIG. 6—Plan of standard Tokatea Hill section at Petote Stream, and Manaia Hill Group in Cadman Stream. Symbols as for Fig. 2; biv—Beesons Island Volcanics; mf—Moehau Formation; thf—Tokatea Hill Formation; blk—black; brn—brown; pyr—pyritised; arg—argillite; zs—siltstone; lam.—laminated; ss—sandstone.

Coarse sandstones and conglomerates are absent, but fine grey sandstones (0·4 mm down to matrix grain size) and grey to brown siltstones are interbedded with dark grey to black, fine siltstones and minor argillites between the typical light grey massive horizons. The sandstones consist of angular to subrounded quartz, lesser plagioclase, and minor orthoclase, rock fragments and accessories, all irregularly scattered throughout a fine (less than 0·01 mm), light coloured sericite, or darker sericite-chlorite matrix. Lithic detritus is virtually absent except at the base of the formation.

The black argillites are very fine-grained rocks consisting of dark skeins of chlorite, clay and pyrite enclosing fine fragments of quartz. In contrast to the light-grey mudstones, a finely laminated texture is present and associated siltstones are brownish. The blue-grey argillites contain a higher proportion of sericite and are associated with blue grey sericitic siltstones. All rocks, but particularly the light grey massive beds and the black argillites, are strongly pyritic.

The area occupied by the formation is considerably less than that mapped by Fraser and Adams (1907), not only because of the attribution of the rhyolites to the Tertiary, but also because a large part of the area between Coromandel and Matawai Stream has been mapped as Moehau Formation on the basis of the petrography and structure of the rocks.

The examination of 50 thin sections and the modal analyses of six coarser representative samples have indicated that, following Pettijohn's (1957) criteria, the Tokatea Hill Formation is a feldspathic greywacke suite (*see* Table 2).

## *Petrography*

### *Quartz*

14% to 35% (average 27%) of detritus. Angular to subangular (AU11728, AU11739), but occasionally rounded (AU11730) or corroded (AU11732). Fracturing, strained undulatory extinction, and secondary overgrowths (AU11730) are rare. Inclusions of apatite and rarely zircon (AU11731).

### *Plagioclase*

1% to 22% (average 12%); subordinate to quartz except near base of formation although grains often large (0·75 mm). Subangular. Normal composition between $An_{24}$ and $An_{35}$ but dusty, rounded untwinned albite ($An_5$) is common (AU11730, AU11740) and basic plagioclase ($An_{70}$) is rare (AU11736). Co-existing fresh albitised, and saussuritised plagioclase is common to all samples. Inclusions of apatite.

### *Orthoclase*

0% to 3%. Uncommon, variable constituent (AU11739, AU11740), occurring as small, rounded and perthitic (AU11735) fragments which stain poorly.

### *Volcanic Rock Fragments*

0% to 10% (average 3%). Very uncommon except at base of formation. Most grains are rounded trachytic-textured andesite/dacite (plagioclase $An_{34}$), but occasional felsophyric "rhyolites" which stain with cobaltrinitrite (AU11731), and rare porphyritic hornblende andesite ($An_{70}$) fragments (AU11736) are present.

### *Other Rock Fragments*

0% to 3%. Fragments are commonly argillite and fine siltstone, or rarely quartzite, chert, and coarsely rhombic calcite with secondary calcite rims (AU11729, AU11730). Plutonic and metamorphic fragments are extremely rare.

TABLE 4—Accessory Minerals of Tokatea Hill Formation
Symbols at head of columns are explained in Key to Table 2, *see* p. 207 (r = rare; x = abundant; v = very abundant)

| | a | b | c | e | g | h | m | p | t | z | A | P | H | F | S |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 3985 | | x | | | r | | x | | | r | r | | | | |
| 3983 | | x | | | r | r | x | r | | | | | | | |
| 11741 | | | | | r | | x | | | r | | | | | |
| 11740 | | x | | x | r | | x | | | x | r | | | | r? |
| 11739 | | x | | | r | | v | | | r | r | | | | |
| 11738 | | | | r | x | | x | | x | v | x | | | | |
| 11737 | | x | | | | | x | | | | | | | | |
| 11736 | | | | r | | | | | | | | r | | | |
| 11733 | | x | | r | | r | x | r | | x | r | | r | | |
| 11732 | | r | r | x | | | x | | | | | | | | |
| 11731 | | | x | | | | | | | | | | | | |
| 11730 | | r | | | | | x | | v | r | | | | | |
| 11729 | | | | | | | r | | | r | | | | | |
| 11728 | | x | | | | | x | | | | | | r | | |

*Accessories* (Table 4)

Characteristically high proportion (0·5% to 2·5%) most of which are micas of which yellowish plates and flakes of pleochroic muscovite (AU11739) predominate over light brown biotite (AU11733; AU11737); dark brown biotite is rare.

Zircons are large and rounded (AU11738, AU11740) but prismatic epidote (AU11732, AU11740) and small colourless to yellowish, rounded to euhedral garnets (AU11738) are rarer.

Ferromagnesians are very rare and include small rounded hornblende (AU11733) or augite (AU3983), and fractured prismatic hypersthene (AU11728, AU11733); actinolite is absent.

Apatite occurs right through the formation but chlorite and pennine are restricted to the base (AU11731–2, AU11736). Sphene occurs as a single rounded fragment (AU11740).

Tourmaline (AU11730, AU11738) occurs as rounded blue-green to pale green or pink prisms up to 0·5 mm in length in scattered clumps.

*Matrix*

34% to 80% (average 55%). Typically light coloured except for occasional streaks of limonitic clay (AU11733) and the matrix of the dark argillites. Consists of scattered grains (AU11728) or streams (AU11730) of quartz surrounded by a dense felt of pale yellow sericite flakes (AU11734) and scattered large plates (AU11729) which together make up 70% of the matrix. The remainder consists of finely divided cherty quartz (AU11730) and the fine green chlorite of the "spotted" rocks (AU11728–9). Less sericitic patches stain with cobaltrinitrite (AU11728, AU11736) and may be unreconstituted potassic clay. Although much pyrite and calcite may be hydrothermal in origin, small lenses and stringers of pyrite parallel to the bedding of the dark argillite and siltstone appear to be detrital.

*Secondary Mineralogy*

Apart from reconstitution of the matrix clays to sericite and chlorite, secondary minerals are restricted to occasional interstitial orthoclase (AU11736), albite (AU11730), and rare, very small (0·05 mm), optically positive, twinned, euhedral ?zeolite (AU11730). Near a base metal-quartz lode in Petote Stream, hydrothermal

alteration has coarsened the matrix sericite and caused sericitisation of detrital feldspars; adularia and epidote are associated with the gangue minerals of the lode (AU11731–2; AU11740–1). Rare tourmaline rosettes (AU11730, AU11742), and pale blue-green needles (0·01 × 0·003 mm) of ?actinolite marginal to quartz grains also occur near Petote Stream (AU11730, AU11737, AU11742).

The strong pyritisation on joint planes and within the rock matrix is undoubtedly of hydrothermal origin, although it may have partly originated by remobilisation of syngenetic detrital pyrite deposited in the dark argillite and siltstone.

## *Origin*

### *Provenance*

The detrital characteristics of Tokatea Hill Formation in ascending the succession from Moehau Formation may be summarised as follows (Fig. 7):

(a) A rapid decrease in volcanic content from little to none at all.
(b) A marked increase in matrix and quartz content.
(c) An increase in stable accessory minerals (micas, zircon, tourmaline, garnet) but a virtual absence of unstable ferro-magnesians except near the base of the formation.
(d) A decrease followed by small increase in orthoclase content.
(e) An irregular but often locally high plagioclase content.
(f) A slight decrease of non-volcanic rock fragments, including the disappearance of almost all but those of possible intraformational origin.
(g) The continued presence of detritus of hydrothermal origin (saussuritised and albitised plagioclase associated with unaltered plagioclase, vein calcite, epidote and pyrite).

These trends can be interpreted as an indication that the relief of the volcanic landmass that was the source of the Moehau Formation had been considerably lowered; the rate of erosion would be much slower and fine silt and clay would form much of the detritus, unstable minerals and rock fragments becoming rare. The supraplutonic volcanic cover may have disappeared altogether so that granodiorites and diorites supplied much of the coarse debris (quartz, plagioclase, some albite, perthitic orthoclase, biotite, muscovite, and zircon). The paucity of chlorite in the quartz-sericite matrix also suggests that the more ferriferous volcanic rocks had little part in the provenance, and that potassic clays derived from alumino silicates only were being deposited. However, it is clear from the albitised and saussuritised plagioclase, the calcite rhombs, and the epidote fragments that hydrothermally altered rocks must have been present in the plutonic basement. Fragments of metamorphic origin are rare or absent, but it is possible that at least part of the muscovite and the rare dark brown biotite could have had this source. The garnets and blue to pink tourmaline could be metamorphic, but could also be derived with large muscovite flakes from pegmatites (Pettijohn 1957).

Figure 7 has been drawn from seven samples located progressively across the Tokatea Hill/Moehau Formations boundary in Petote Stream together with the average of three samples from the underlying conglomerate band in Matawai Stream (AU11675, AU11716, AU11717). The smoothed curves may be interpreted as generally indicating a progressive lowering of relief, but the sharp change in feldspar content suggests a sudden transition from

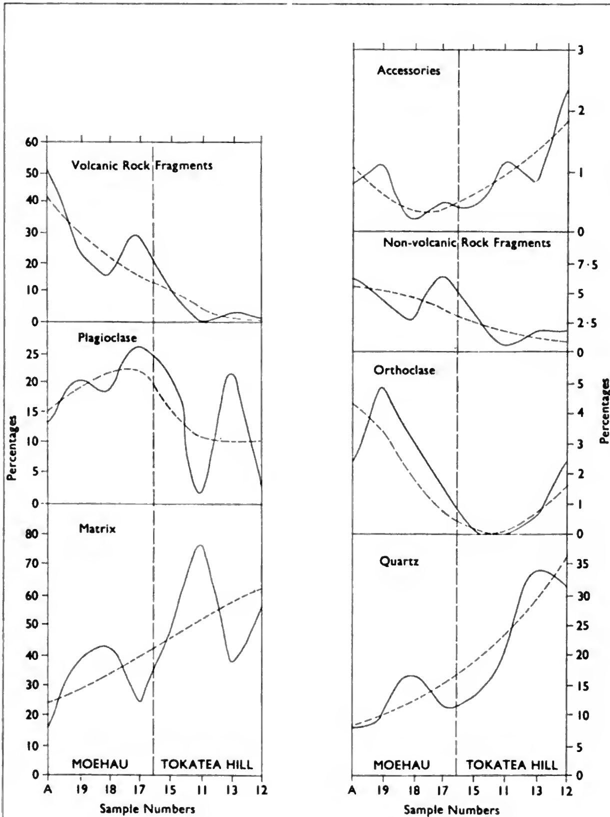


FIG. 7—Composite variation diagram of progressive modal changes throughout a stratigraphic thickness of 375 m either side of the Moehau–Tokatea Hill formation boundary (dashed vertical lines). Sample numbers refer to Table 2; A = average of samples 4, 8, and 9 of Table 2. Solid curves join actual points; dotted curves are smoothed, generalised trends.

a volcanic to plutonic source. In addition, the general increase in stable detritus and matrix with the concommitant loss of less resistant clasts indicate a progressive increase in maturity of the sediments.

### *Sedimentational Features*

The sedimentary features of each lithology are characteristically unique. The light-grey horizons tend to be massive and show only rare traces of internal bedding in the form of very faint, fine laminations which are contorted into intraformational convolutions. These massive beds grade into light blue-grey siltstone and mudstone which are well bedded but slightly boudinaged. Lenses of breccia (2 m × 50 cm) which parallel the bedding consist of angular fragments of the light-grey beds embedded in an almost white, heavily pyritised, sericite-rich matrix. Rarely, the light blue-grey siltstones and mudstones have poorly developed, fine micro-current bedding.

The black argillites and dark siltstones are finely and evenly laminated but are often contorted into large irregular slumped blocks bounded by shears. Load casts, small flame structures, and scour casts are common, but rarely exceed 1 cm in height. Although beds may be 10 cm thick, the individual laminae range in thickness from less than 1 mm in the argillites up to 10 mm in the siltstones. Grading is not uncommon both within individual beds, and progressively from one thin bed or lamina to the next, in which case each is separated by a black argillite layer. Breccia dikes originating in the underlying light-grey beds have been squeezed up joint planes cutting the darker sediments.

### *Environment of Deposition*

The Tokatea Hill Formation may be unique amongst described rocks of the New Zealand Geosyncline, in that it is a unit of feldspathic greywackes derived from intermediate to acid calc-alkaline plutonic rocks, with low grade metamorphic and volcanic detritus in the basal members. Hence it is very different from Moehau Formation and Wellman's (1952) shelf or Hokonui Facies (calc-alkaline, lithic volcanic), and from the Alpine Facies or deep water axial turbidites (feldspathic; acid plutonic–high rank metamorphic) of Reed (1957).

The sedimentary features are similar to those of both the Hokonui and Alpine facies. The angular (micro-breccia) texture and freshness of the detritus implies rapid or short transport whereas the poor sorting ("poured in, mud-slurry") implies rapid deposition. The light grey beds near the top of the formation appear to have a higher grade of sedimentary maturity in that they have a very high quartz/feldspar ratio; however, the proportion of matrix is also rather high.

It thus appears likely that the lower parts of the formation are redeposited turbidites extending from a slope to deep water environment (grading, scour casts, load casts, laminated and convolute bedding, flame structures) although the sedimentational features are less marked than those of the axial Alpine Facies. The upper parts of the formation may then be the product of quieter, stable deposition in fairly deep water (greater maturity, little bedding, massive beds, finer grade, no fossils).

### *Structure*

The structure of the Tokatea Hill Formation is essentially simple. The strike varies between north-west and north, although there are slumped, irregular horizons; the dip is generally very steep to the west (65° to 80°). Scour and load casts, and grading and current bedding in the rocks of Petote Stream, indicate that the strata young to the west. In Cadman Stream immediately to the north, Moehau Formation crops out both west and east of steeply dipping Tokatea Hill Formation and young to the east and west respectively, thus suggesting that the Tokatea Hill sediments lie in the core of a syncline. In the headwaters of Matawai Stream, the beds again dip steeply west under overturned Moehau Formation; poor grading suggests Tokatea Hill Formation is also overturned, i.e., it youngs to the east. In Tiki Stream Moehau sediments to the west of Tokatea Hill Formation are also overturned and dip to the west.

Thus Cadman Stream would seem to expose the whole of a syncline of Tokatea Hill Formation although outcrops are not very good; the Petote Stream standard section would then be the eastern limb, and the Matawai Stream exposures would be the western limb of the syncline (Fig. 2, 3).

Well developed jointing is not common but when present also confirms a north-west to north-north-west major fold axis.

## SUMMARY OF INTERNAL RELATIONS

### *Petrography*

In all sections where a boundary between the two formations is present, the rocks at the interface are gradational, the typical rocks being defined as distinctly lithic-volcanic and distinctly feldspathic greywackes respectively. The most obvious example of gradation (Fig. 7, 8) is the volcanic rock fragment content which becomes less towards the top of Moehau Formation, is low in basal Tokatea Hill Formation, and virtually absent in the upper part of this formation. A reverse gradation is indicated by the crystalline detritus and matrix, the latter being characteristically sericitic in Tokatea Hill Formation; sericite occurs in Moehau Formation only near the formation boundary. Figure 8 indicates steadily varying conditions except for feldspar which, instead, shows a sudden marked decrease in Tokatea Hill Formation. This is perhaps due to the lack of both orthoclase and plagioclase of volcanic origin, the detritus being mainly derived from plutonic rocks.

Triangular plots of quartz-feldspar-rock fragments, quartz-feldspar-matrix, and quartz-matrix-unstable clasts (feldspar and rock fragments) clearly indicate the distinction between the lithic volcanic Moehau Formation and the feldspathic Tokatea Hill rocks (Fig. 9). In addition, the samples from stratigraphically either side of, and through the formation boundary occupy a median position on the graphs, and plot progressively from one side of the graph to the other. The sample grouping designated as transitional includes those rocks collected from near the top of Moehau Formation

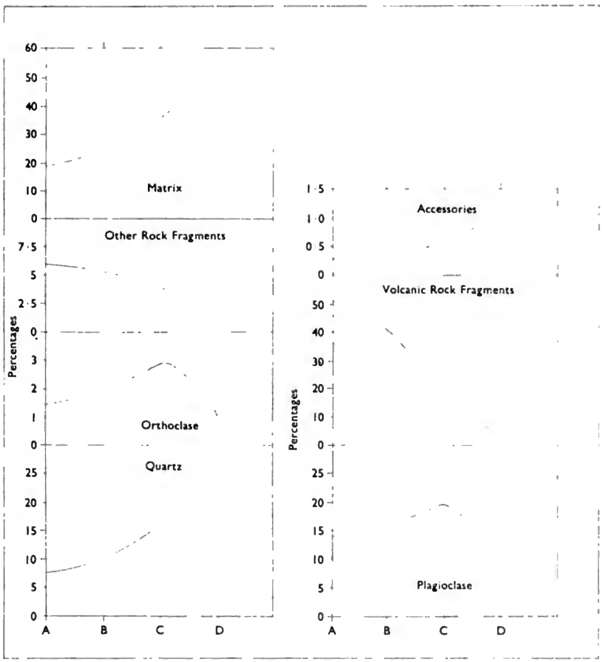


FIG. 8—Summary modal variations of Manaia Hill Group.

A typical lithic volcanic greywacke suite;
B average of all Moehau Formation;
C rocks gradational between lithic volcanic and feldspathic greywacke suite;
D typical feldspathic greywacke suite (Tokatea Hill Formation).

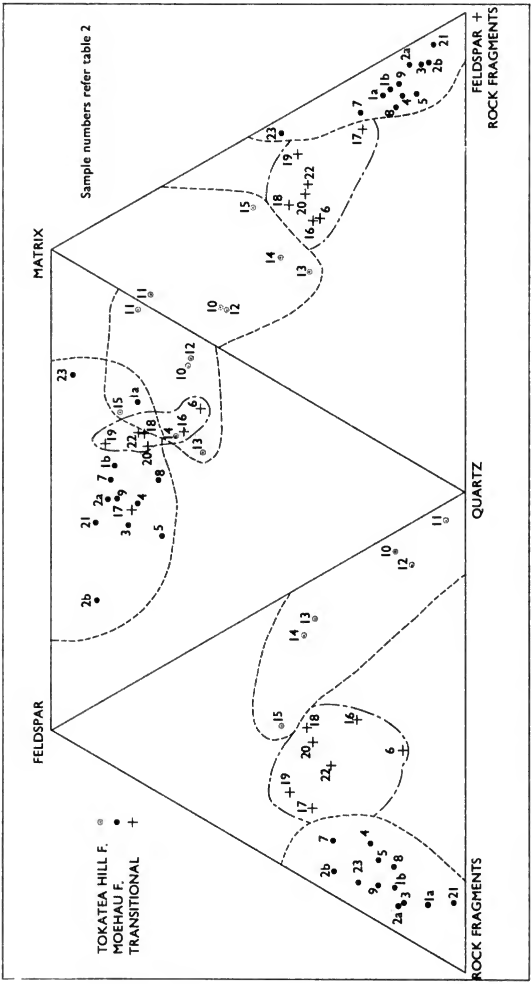


FIG. 9—Triangular modal variation diagrams of Manaia Hill Group; sample numbers refer to Table 2.

which are poorer in rock fragments and a little richer in feldspar than the bulk of that formation; sample 6 is included because of an inordinately high quartz component although in gross detrital content it is obviously a lithic volcanic greywacke of Moehau Formation.

### *Structure and Rock Unit Distribution*

On the basis of dips, younging of beds, and conformity of strikes, the Manaia Hill Group at Coromandel is considered to form an isoclinal, somewhat overturned anticline and adjoining syncline, with the Tokatea Hill Formation occupying the trough of the syncline, and being the younger of the two formations rather than the older as designated by earlier workers. Without the aid of younging directions, however, the sequence would certainly have appeared to earlier workers to have been the reverse. The section through the Tokatea Tunnel described by Maclaren (1900) thus represents not one formation of "felsites" sandwiched between slates and greywackes, but a synclinal section of two formations. Similarly the Tokatea-Tiki range does not consist of Tokatea Hill rocks alone (as mapped by Fraser and Adams (1907), but instead includes interfolded Moehau and Tokatea Hill Formations.

## ACKNOWLEDGMENTS

This paper is part of the results of Ph.D research done at University of Auckland under Professor A. R. Lillie and Professor R. N. Brothers. The work was financed by field and research grants allocated by the University Grants Committee, and by study leave granted by N.Z. State Services Commission and N.Z. Geological Survey.

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# NEW RECORDS OF THE ELASMOBRANCH *C. MEGALODON* (AGASSIZ) AND A REVIEW OF THE GENUS *CARCHARODON* IN THE NEW ZEALAND FOSSIL RECORD

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ABSTRACT

New Zealand specimens of fossil shark teeth referable to *Carcharodon megalodon* (Agassiz) are discussed, including the first examples from the North Island. Previously published records are reviewed and revised, and the stratigraphic distribution of this species discussed. The New Zealand records of *C. megalodon,* which begin in the Lower Oligocene, provide some of the earliest world records for this species. Stratigraphic distribution of the two related species, *C. auriculatus* Blainville, and *C. carcharias* (Linnaeus) is also examined.

## INTRODUCTION

*Carcharodon megalodon* (Agassiz) is the largest Elasmobranch known to have lived during the Tertiary. Because of their huge size, teeth of this species are among the best known of all fossil shark teeth. It is estimated that these sharks reached 30 metres in length with a gape of 2 metres. Records of this cosmopolitan species come from the Tertiary sediments of most countries of the world, but as collections from the *Challenger* and *Albatross* deep-sea dredging expeditions suggested, the species may have survived until late Pliocene–Pleistocene times in the Pacific.

New Zealand fossil elasmobranch teeth were first studied by Davis (1888). Later Chapman (1918) revised Davis' work (repeating his illustrations) and the results were published as *N.Z. Geological Survey Paleontological Bulletin* 7. When dealing with *Carcharodon megalodon* (Agassiz), Chapman (1918) had only the same three specimens available to him (all from South Island localities) as did Davis, so presented no new information. As no further records of this species have appeared for over 80 years since Davis' work it has been generally concluded that *C. megalodon* was an extremely rare shark in the New Zealand Tertiary. In this paper, the localities and ages of Davis' and Chapman's three specimens are reviewed and records of 17 specimens that have recently been located are given, including the first specimens from the North Island. The more outstanding examples are figured. This extra material has been the result of extensive inquiries and examination of specimens in collections of New Zealand universities and museums. Specimens here identified as *C. megalodon* represent material that is sufficiently well preserved to be confidently ascribed to this species. Damaged or incomplete specimens that could be doubtfully placed in this species have been omitted.

The specimens described in this paper are grouped into North and South Island records and arranged geographically according to NZMS 1 sheet districts. As most specimens are from old museum collections it has been necessary to comment on their locality details in order to establish their stratigraphic position. Ages of Lower Miocene specimens which are cited as Altonian are given in terms of the new Altonian Stage and stage revision proposed by Scott (1972). Dimensions of specimens exclude the root as this is usually damaged or incomplete. Measurements of teeth are taken along three approximately perpendicular directions. Crown height is taken parallel to the outer face from apex to base, (base being taken as a line joining the lowest extent of crown enamel at the sides of the tooth), perpendicular to the base; width is the maximum transverse measurement across the crown at lowest level of root/crown junction; thickness is taken in a medial position on the crown normal to the outer face above the root/crown junction.

## Superclass PISCES

## Class CHONDRICHTHYES

## Subclass ELASMOBRANCHII

## Infraclass OSTEODONTA

## Order SELACHII

## Family ISURIDAE

### Genus Carcharodon Müller and Henle, 1841

### Carcharodon megalodon (Agassiz)*

1835 *Carcharias megalodon* Agassiz, Feuilleton 72, *fide* Sherborn, 1928, Index Animalium 13: 3941.*

1837 *Carcharias megalodon* Agassiz; Charlesworth, Mag. Nat. Hist. n.s. 1: 225, fig. 24.

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*Confusion exists in the literature as to the correct authorship of the species *C. megalodon*. This may have arisen through authors referring to Woodward's (1889, p. 415) synonymy for the species where Charlesworth's (1837) use of the name *Carcharias megalodon* to label a fossil tooth he illustrated appeared to establish a date priority over Agassiz whose detailed study of fossil fish was not published until 1843. However, an examination of Charlesworth's (1837) paper shows that he identifies his figure (fig. 24, p. 226) as "*Carchàrias megâlodon* Agass.", confirming that the name he used definitely belonged to Agassiz and was available prior to 1837. Agassiz (1843, p. 247) placed *Carcharias megalodon* in the genus *Carcharodon* and included in his synonymy only a single reference to an earlier identification of "*Carcharias megalodon* Agassiz" made by P. de M. G. Egerton in one of the latter's catalogues, either the 1836, 1837 or 1841 work (British Museum 1904, p. 513). Sherborn (1928, p. 3941), however, lists a much earlier use of "*Carcharias megalodon* Agassiz" by Agassiz in 1835 in "Feuilleton, 72", which was one of a series of separately published articles (Sherborn, 1922, p. xv) issued between 1834–43 by Agassiz on "Poissons Fossiles". Although Charlesworth (1837) may have provided the first identified illustration of *C. megalodon,* priority for the authorship of this species is definitely attributable to Agassiz both on account of his 1835 usage as well as his 1843 fully illustrated definitive description of the species.

1843 *Carcharodon megalodon* Agassiz, Récherches sur les Poissons Fossiles Vol. 3, p. 247, pl. 29.
1888 *Carcharodon megalodon* Agassiz; Davis, Sci. Trans. R. Dublin Soc. 4 (Ser. 2): 12, pl. 2, fig 1–3.
1889 *Carcharodon megalodon* Agassiz; Woodward, Cat. Fossil Fishes in British Mus. (Nat. Hist.), Pt. 1, p. 415.
1904 *Carcharodon megalodon* (Charlesworth), Eastman, Maryland Geol. Surv. Miocene, p. 82, pl. 31, fig. 1–4.
1918 *Carcharodon megalodon* Agassiz; Chapman, N.Z. geol. Surv. Paleont. Bull. 7: 19, pl. 2, fig. 1–3.
1921 *Carcharodon megalodon* (Charlesworth); Ishiwara, Sci. Rep. Tohoku Imp. Univ. Ser. 2 (Geol.) 5: 65, pl 10, fig. 33; pl 11, fig. 1–8; pl. 12, fig. 1–2.
1926 *Carcharodon megalodon* Agassiz; Leriche, Mém. Mus. roy. Hist. nat. Belg., Ser. 1, vol. 32: 412, pl. 35, 36.
1960 *Procarcharodon megalodon* (Agassiz); Casier, Ann. Mus. Roy. Congo Belge A, Ser. 3, Tome 1, fasc. 2: 13.

[Abbreviated synonymy only, listing principal New Zealand references, main historical references and works that illustrate the species. Extensive synonymy for the species is contained in Woodward, Eastman, Chapman and Leriche.]

DIAGNOSIS: Very large triangular, robust teeth. Crown broad, rapidly widening towards base. Outer coronal face flat; inner, convex towards apex and concave towards base. Margins finely serrated. Root large, not extending beyond width of crown base. No distinct lateral cusps.

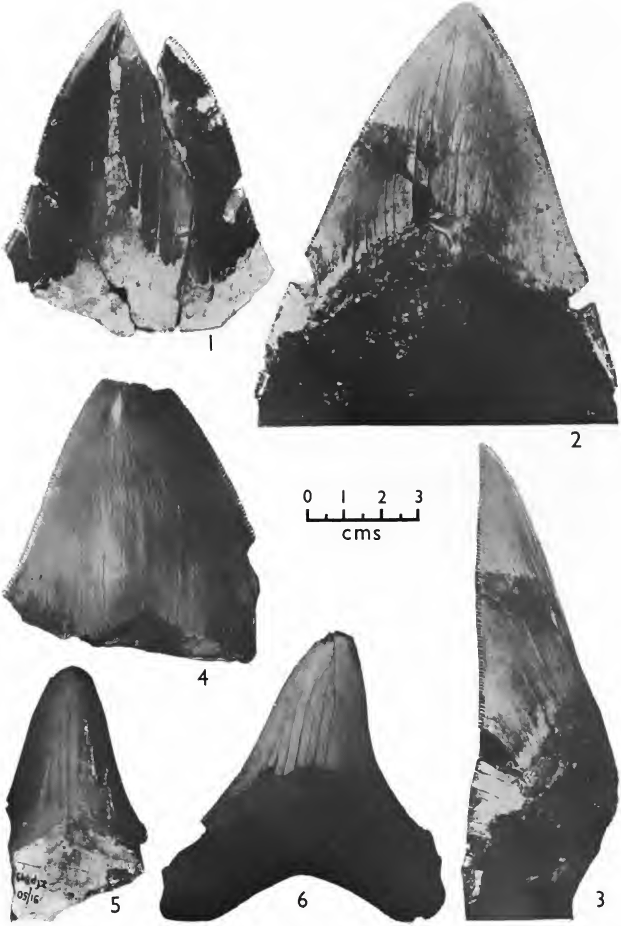
## NORTH ISLAND SPECIMENS

1. *Ihungia Stream, N Gisborne* N80. Broken basal portion of an anterior tooth in the Gisborne Museum. DIMENSIONS: width 67 mm; thickness 23 mm. The specimen was collected from the bed of the Ihungia Stream, 1 mile upstream from the Ihungia-Waitahaia road bridge. Although broken, the remainder of the tooth is in good condition suggesting that it was collected from a locally derived boulder. Age of the adjacent beds (Kingma 1965) is Altonian (Lower Miocene).

2. *Patutahi Quarry, Gisborne* N97. Anterior tooth (in matrix) with damaged apex and lower crown margins on loan to Gisborne Museum. (Cast held at N.Z. Geological Survey.) DIMENSIONS: Crown height (est.) 80 mm; width 63 mm; thickness (est.) 10 mm. The specimen was obtained from rock excavated from one of the two quarry faces being worked at the Patutahi Council Quarry. The matrix is a sandy shelly limestone, mapped (Kingma 1964) as Opoitian–Mangapanian. An Opoitian (Lower Pliocene) age however is most likely for this specimen.

3. *Hangaroa, N Hawke's Bay* N97. (Fig. 1). Large, damaged, anterior tooth in Dominion Museum Elasmobranch collection (No. 5268). Apex of crown damaged and root missing, and lower crown margins broken. DIMENSIONS: Crown height (est.) 78 mm; width (est.) 80 mm; thickness 22 mm. Locality details with the specimen are "Hangaroa, Hawkes Bay". The rocks in the

FIG. 1–6—*Carcharodon megalodon* (Agassiz)
Inner coronal views except Fig. 3 (side).
1—Hangaroa. Dominion Museum No. 5268.
2, 3—Pipiriki, Wanganui Museum Drew Collection No. 4373.
4—Wanganui. British Museum (Nat. Hist.) No. P. 5809.
5—Hawarden. Canterbury Museum No. zfp. 849.
6—Trelissic Basin, British Museum (Nat. Hist.) No. P. 2309.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immediate vicinity of Hangaroa Township are sandstones and siltstones mapped as Tongaporutuan (Upper Miocene) (Marwick 1965, map). As the specimen was collected broken into 6 separate pieces is was probably extracted from a hard concretionary band.

4. *Pipiriki, NW Wellington* N121. (Fig. 2, 3). Large, well-preserved, anterior tooth with damaged root; in the Wanganui Museum. DIMENSIONS: Crown height 118 mm; width 97 mm; thickness 29 mm. This specimen was obtained by the Wanganui Museum in 1895 as part of a purchase of natural history specimens in the collection of Mr S. H. Drew, of Wanganui. The registration number on the specimen (No. 4373) is a record number from the Drew collection, but the Museum possesses no catalogue details apart from the fact that the specimen was originally collected from "Pipiriki" (a small settlement some 59 miles up the Wanganui River). The matrix associated with the tooth is a grey-brown, slightly micaceous sandstone with occasional fine carbonaceous specks and, according to Mr D. S. Ker (N.Z. Geological Survey, pers. comm.), resembles lithologies found in the Matemateaonga Sandstone (Hay 1967) north of Pipiriki. This formation in the Pipiriki area has been mapped as Opoitian (Lower Pliocene).

5. *Wanganui, NW Wellington* ?N136–137. (Fig. 4). The first specimen known to have been collected from the North Island has been in the possession of the British Museum (Natural History) since last century. Chapman (1918) did not mention this tooth although it had been listed by Woodward (1889, p. 420, specimen P.5809). The locality details quoted by Woodward are confused, making it uncertain whether it came from New Zealand or Australia, but a recent examination of the records (by Mrs C. B. Welch, Fossil Fish Section, British Museum) has enabled more satisfactory details to be obtained from the original locality label. These are: "Probably from the Upper Miocene beds Older Wanganui Series of N.Z. Geol. Surv. From between Wanganui and N. Plymouth. Signed (?) M. Hester". [Probably signature of J. Hector, Director of the N.Z. Geological Survey and Colonial Museum, 1865–1904.] The specimen is a large, imperfect crown of an anterior tooth broken at apex, crown base and along one lateral margin. DIMENSIONS (from remaining specimen): Crown height 69 mm; breadth 70 mm; thickness 25 mm. The exposed beds along the Wanganui–Taranaki coastline range in age from Opoitian to Castlecliffian. The age of the tooth is therefore Lower Pliocene to Middle Pleistocene.

## SOUTH ISLAND SPECIMENS

6. *Cape Foulwind, Westland* S23. This large, well-preserved specimen of an anterior lateral tooth, in the collection of the N.Z.. Geological Survey, was figured by Davis (1888) and Chapman (1918) in plate 2, fig. 1a–c of their works and later by Stevens (1966, pl. 42A, fig. 7), Maughan (1970, p. 54) and Keyes (1971, fig. 1). Both Davis and Chapman cited "Cape Foulwind, Westport, Nelson" as the locality. Morgan (*in* Chapman, p. 35) was uncertain whether the specimen came from above or below the Cobden Limestone at Cape Foulwind. This absence of essential stratigraphic data led Stevens (1966) into ascribing a "Lower Tertiary" age to the specimen. A

search through Suter (1921) however has revealed (p. 35) that under Geological Survey locality No. 823 for "Cape Foulwind, Westport" the following statement is made which ties this tooth to a definite locality: "McKay in one of his MS. lists assigns this locality to the Pareora Series, and mentions that the collection (three specimens) contains a very large tooth described by Davis (*Carcharodon megalodon*)". The only fossil listed by Suter for this locality is a gastropod *Miomelon corrugata* (Hutton). Dr A. G. Beu has examined this specimen and has reidentified it as *Alcithoe* cf. *gravicostata* King which ranges from Tongaporutuan to Kapitean. This specimen of *C. megalodon* is now assigned to a definite Upper Miocene age.

7. *Hawarden, N Canterbury* S61. (Fig. 5). Worn anterior tooth broken at crown apex and lower lateral margins. Broad outline and relative thinness of the specimen suggests that it belongs to this species. DIMENSIONS: Crown height (est. 55 mm); width (est. > 50 mm); thickness 14 mm. Specimen in the Canterbury Museum (No. zfp. 849), collected "from limestone" at Hawarden. The limestone referred to is most likely to be the Amberley Limestone which occurs to the south of Hawarden township. Age is therefore Whaingaroan (Lower Oligocene).

8. *Trelissic Basin, Canterbury* S66. (Fig. 6). A further specimen of an anterior tooth from New Zealand was recorded by Woodward (1889, p. 420) in the British Museum collection. The specimen (P. 2309) was received by the British Museum (Natural History) in 1876 and the published locality details gave "Trelissic Formation, New Zealand". Mrs C. B. Welch of the British Museum kindly re-examined this specimen and gives the following details (letter 5 June 1970): "The specimen P. 2309 is a *Procarcharodon megalodon* (Agassiz) complete with root; overall length 70 mm; breadth 50 mm (at base of crown). All the characters of *P. megalodon* are present although the specimen is comparatively small". The locality details from the original Museum registration book were also checked and give "Trelissic Basin, Ototara Series, Canterbury New Zealand". The old term "Ototara Series" which included rocks now placed in several formations (Gage, 1959, p. 303) has been discontinued, but equivalent rocks are presently included in the Porter Group (Gage, 1970, Table 1, p. 510) of Whaingaroan–Duntroonian (Lower–Middle Oligocene) age.

9. *Bobys Stream, Waipara, N Canterbury* S68. This specimen was figured by Davis (1888) and Chapman (1918) in plate 2, fig. 3. Davis (p. 13) incorrectly quoted the locality as "Cobley's Creek", but Chapman corrected this to Bobys. Morgan (*in* Chapman, p. 37) identified the matrix with the specimen as "Weka Pass Stone". The matrix attached to the tooth is typical of the Weka Pass Stone, being a hard whitish glauconitic calcareous sandstone. Although the tooth is labelled "Boby's Creek", Thomson (1920, p. 355) claimed that there are no outcrops of Weka Pass Stone in Bobys Stream and that "the specimen may have come from the slopes of Mount Brown" to the south-east of this stream. With the tooth is its original Geological Survey label dating from last century which reads "near Bobby's Creek" thus indicating a locality away from Bobys Stream. As a Geological Survey specimen it probably was collected from a numbered locality. Either collection GS 74 or GS 724, made last century from "Weka Pass calcareous greensands" (by A. McKay, 1874 and later) and "Lower calcareous Band, Mt

Brown" (J. Park, 1887) could be the source of the tooth. The specimen is therefore of Duntroonian–Waitakian (Middle-Upper Oligocene) age.

10. *Weka Pass, N Canterbury* S68. Specimen of an anterior lateral tooth, figured by Davis (1888) as plate 2, fig. 2 and also repeated by Chapman (1918), in the Canterbury Museum (No. zfp 122). Davis and Chapman listed the locality as "Weka Pass" and Morgan (*in* Chapman 1918, p. 37) could not elucidate further on this. As the specimen does not have any adhering matrix a particular formation cannot be suggested. However there is a strong likelihood that this is the specimen mentioned by Hutton (1888, p. 262) that he collected from "Greta Beds" at Weka Pass. This would make the age Opoitian–Mangapanian (Pliocene).

11. *Teviotdale, Waipara, N Canterbury* S68. (Fig. 7). Anterior lateral tooth with broken root, from upper jaw. DIMENSIONS: Crown height 81·5 mm; width (est.) 77 mm; thickness 17·5 mm. The specimen, in the Canterbury Museum (No. zfp 850), was collected "on Teviotdale farm, Waipara, high up on seaward side". The collecting spot was probably in siltstone beds on the coastal range mapped as Southland Series (Wilson 1963). The yellow-brown colour is very reminiscent of teeth from Mount Brown beds (Pareora Series) which may underlie this area. The specimen is therefore thought to be of Altonian (Lower Miocene) age.

12. *Waimakariri River (?), N Canterbury* S74 (?). Half an upper anterior tooth with broken crown apex, in Auckland University. Overall crown height from remaining specimen 90 mm; thickness 24 mm. The specimen is reputed to be from "Calcareous sandstone, Otarama Gorge, Waimakiriri River, North Canterbury" (S74/f634), but the preservation of fossils in these beds (of Paleocene–Cretaceous age) is considerably different from that of the tooth. Small fragments of yellow-brown limestone adhering to the broken surface of the tooth confirm that it must have come from younger beds, possibly from the Thomas Formation of the Broken River area to the north. A Duntroonian–Waitakian (Middle–Upper Oligocene) age is likely.

13. *Burnt Hill, Oxford, N Canterbury* S75. Broken fragment of crown apex of an anterior tooth, probably from an upper jaw, 27 mm long, in N.Z. Geological Survey macrofossil collection GS 3534 from "Burnt Hill, Oxford" (S75/f499). Extreme broadness of this fragment allows it to be placed with confidence in this species. At Burnt Hill, between the main Waiauan shellbed and underlying volcanics there is a greensand horizon containing shark teeth, bones and phosphatic nodules (Bed 2 of Speight 1928, p. 422) from which this collection was obtained. The greensand suggested a Landon Series age, and a brachiopod specimen obtained in a later re-collection (S75/f516) of this bed was identified as *Waiparia (?) elliptica* (Thomson) by Prof. R. S. Allan (letter from Mr D. R. Gregg, 27 September 1962) which indicates that the horizon is Duntroonian (Middle Oligocene) in age.

14. *"Curiosity Shop", Rakaia River, Canterbury* S82. (Fig. 8). Upper portion of crown and part of one side of a broken and worn anterior tooth, from an upper jaw. Length of remaining specimen 56 mm. Specimen in the Canterbury Museum (No. zfp 868). Locality is "probably Curiosity Shop, Rakaia River". The Waitakian greensand lying above the calcareous sandstone at "Curiosity Shop" (a locality situated on the NE bank of Rakaia River, 5 miles below the gorge) has yielded a rich shark fauna in the past (Morgan *in* Chapman


FIG. 7–12—*Carcharodon megalodon* (Agassiz)
Inner coronal views except Fig. 8 (outer) and Fig. 11 (side).
7—Teviotdale. Canterbury Museum No. zfp. 850.
8—Curiosity Shop. Canterbury Museum. No. zfp. 868.
9—Kakanui. University of Otago No. OU 9062.
10, 11—Knapdale. University of Otago No. OU 10768.
12—Milburn. Otago Museum No. C.34.2

1918, p. 38) so is likely to be the source of this tooth. However, the underlying calcareous sandstone (Duntroonian) cannot be excluded as the source of the fossil so its age is best regarded as Duntroonian–Waitakian (Middle-Upper Oligocene).

15. *Naseby, N Otago* S135. Small anterior lateral tooth with broken root in collection of University of Otago (No. OU 9055). DIMENSIONS: Crown height 37 mm; width 24 mm; thickness 10 mm. Although small, the specimen possesses the characters of *C. megalodon*. It was collected from the outlet channel just north of the Government dam 1·5 miles west of Naseby, Otago (S135/f562) and is Duntroonian (Middle Oligocene).

16. *All Day Bay, N Otago* S136. Posterior lateral tooth in Geological Survey collection (No. GS 9688). DIMENSIONS: Crown height 45 mm; width 44 mm; thickness 9·5 mm. The specimen is identical with one figured by Leriche (1957, pl. 3, fig. 3). The tooth was collected from the Gee Greensand immediately above the bored surface of the McDonald Limestone at the north end of All Day Bay, N Otago (S136/f1061), and is Waitakian (Upper Oligocene) in age.

17. *McDonald Limestone Quarry, Kakanui, N Otago* S136. (Fig. 9). Anterior lateral tooth with damaged crown apex and lower crown and root margins. DIMENSIONS: Crown height (est.) 47 mm; width (est.) 40 mm; thickness 12 mm. Repository, University of Otago (No. OU 9062). The hard, creamy-white limestone exposed in the McDonald limeworks quarry (1 mile north of Kakanui) is the Flat Top Limestone member of the McDonald Limestone (Gage, 1957, p. 44) of Whaingaroan (Lower Oligocene) age.

18. *Oamaru, N Otago* S136. A large, worn, anterior tooth with crown apex, lower crown margins, and root damaged. DIMENSIONS: Crown height (est.) 75 mm; width 65 mm; thickness 21 mm. This Geological Survey specimen is unlocalised, but Morgan (*in* Chapman 1918, p. 39) suggested that it could possibly have come from the Oamaru District. A distinct possibility would be one of the Oamaru limestone formations of Arnold to Landon Series (Upper Eocene to Oligocene) age.

19. *Waimumu, Southland* S169. (Fig. 10, 11). Very large, broad, upper anterior tooth with broken margin and incomplete root. DIMENSIONS: Crown height 120 mm; width (est.) 117 mm; thickness 22 mm. It must rank as one of the largest specimens of the species recorded. The specimen (No. OU 10768) is in the University of Otago and is recorded as coming from "Waimumu Quarry". Matrix adhering to the base of the crown and root consists of a coarse brown to grey sandstone (which bears the imprint of a fossil *Venericardia*), with quartz pebbles. Thus the specimen is from the Chatton Marine Formation which is worked at Waimumu Quarry for agricultural marl (Wood 1956, p. 14, 117) dated as Duntroonian (Middle Oligocene).

20. *Milburn, Otago* S172. (Fig. 12). Anterior tooth with broken lower crown and lateral root margins. DIMENSIONS: Crown height 50 mm; width (est.) 47 mm; thickness 14 mm. This Otago Museum specimen (No. C.34.2), labelled "Milburn", was almost certainly collected from the Milburn Limestone (where shark teeth occur) and is thus Waitakian (Upper Oligocene) in age.

## DISCUSSION

Although the number of identified specimens of *C. megalodon* from New Zealand has now been increased, the total available, in spite of over 100 years of collecting, is relatively low, and teeth of this species are not as abundant in collections as those of the closely related species *C. auriculatus*.

The stratigraphic distribution of specimens of *C. megalodon* in New Zealand is given in Table 1. The earliest records are from the Whaingaroan (Lower Oligocene). Chapman and Pritchard (1904, p. 293) list a record of *C. megalodon* from the New Zealand Upper Cretaceous but it is not clear from where they obtained their information. The Landon Series (Oligocene) provides most of the records of this species (and other sharks) in New Zealand. This period of time saw the maximum marine transgression of the Tertiary (Fleming 1962, p. 73), and widespread glauconite and limestone deposition in an environment obviously favourable to preservation of sharks' teeth (Keyes 1971). The Oligocene records of *C. megalodon* are important, as the earliest world records for this species (Leriche 1936, table p. 766) appear to be Lower Miocene. The absence of records for the Southland Series may or may not be due to collection failure. Since only 20 specimens of *C. megalodon* are known in the New Zealand fossil record, they are unlikely to represent its total stratigraphic range, but as there are also no records of *C. auriculatus* for this interval it seems possible that the genus *Carcharodon* was absent from the New Zealand area during part of the Miocene. Both species are known from the Middle Miocene of Australia (Pledge 1967, p. 154) and Patagonia (Ameghino 1906, p. 502), so *Carcharodon* was obviously present in the Southern Ocean in the Miocene, when it was abundant in most other areas of the world (Leriche 1936, table p. 766). Like New Zealand, most other regions of the world have produced Pliocene records. Pleistocene records are more restricted, but include those from California (Leriche 1936, p. 766) and from deep sea dredgings in the Pacific (Murray and Renard 1891, pl. 5, fig. 1, 2, 5; Agassiz 1902, Stations 2, 13, 173, p. 70–1, 73–4, identified by Leriche 1936, p. 766). Thus the species may have been restricted to the Pacific during the Pleistocene prior to its extinction.

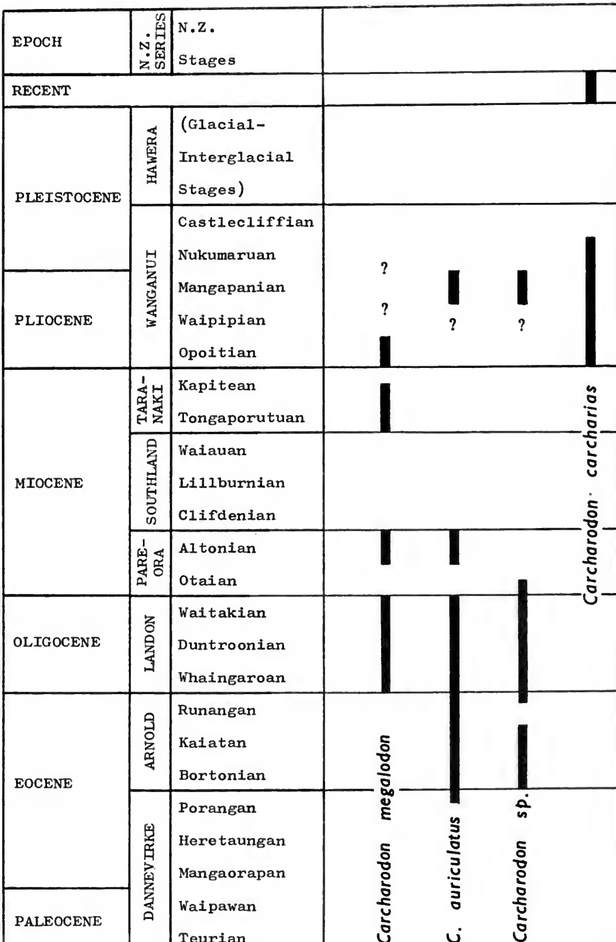
## RELATED SPECIES

In addition to *C. megalodon* there are two further species of the genus *Carcharodon* recorded in New Zealand: *C. auriculatus* (Blainville), and *C. carcharias* (Linnaeus) (= *C. rondeletii* Müller & Henle), the white shark of the present seas.

### *C. auriculatus* (Blainville)

The stratigraphic distribution of this species in New Zealand is recorded in Table 1, prepared from an examination of all identifiable specimens obtainable from Geological Survey, university, museum and private collections, and revision of the material referred to by Chapman (1918, p. 19). The fossil record follows very closely that of *C. megalodon*, by being best represented in Oligocene and Lower Miocene periods, but differs by

TABLE 1—Stratigraphic Distribution of the Genus *Carcharodon* in the New Zealand Cenozoic



having its earliest records in the Middle and Upper Eocene. The record Chapman (1918, p. 19) attributed to the Cretaceous was based on a specimen obtained from the Amuri Limestone at Haumuri Bluffs (N.Z. Geological Survey loc. "G.S. 12"), then believed to be Cretaceous but now known to be younger (Dannevirke–Arnold age). A matrix sample from this specimen was examined for foraminifera and was found to give an age of Porangan–Bortonian (Middle Eocene) (Dr P. N. Webb, pers. comm.) This tooth still provides the earliest record of *Carcharodon* in New Zealand, and corresponds to the earliest records for the genus in the Northern Hemisphere.

Overseas *C. auriculatus* is recorded from the Middle and Upper Eocene of North America, Europe and Africa (Avnimelech 1959; Casier 1960, p. 13), the Oligocene of Europe (Leriche 1910, p. 291) and Patagonia (Ameghino 1906, p. 181) and the Miocene of Australia (Chapman and Pritchard 1904, p. 290; Chapman and Cudmore 1924, p. 134; Pledge 1967, p. 154).

Table 1 also records the stratigraphic range of all the numerous fragmentary specimens of *Carcharodon* that could not confidently be identified to either *C. megalodon* or *C. auriculatus.* This material taken in conjuction with the record for each of the two species (plus that for *C. carcharias*), gives a complete record for the genus in New Zealand, considerably extending the range previously cited (Romer 1966, p. 350) back in time.

*C. carcharias* (Linnaeus)

The fossil record of this species in New Zealand (Table 1) is based on all available specimens from Geological Survey, University, Museum and private collections. It shows a range from Lower Pliocene through to Lower Pleistocene times. This is in close agreement with its fossil record in Victoria (Chapman and Cudmore 1924, p. 136) and South Australia (Pledge 1967, p. 155, age corrected to Pliocene by author, *not* Miocene; *see* p. 158), but it is recorded as well in the Miocene of South America and the northern hemisphere (Leriche 1936, table p. 768).

## CARCHARODON PHYLOGENY

Casier (1960) revised the genus *Carcharodon,* retaining *C. carcharias* (= *C. rondeletti*) as a monospecific genus, and erecting a new genus *Procarcharodon* to include *C. auriculatus, C.. megalodon* and other related species. In this paper the broader genus has been retained and the validity of *Procarcharodon* left an open question.

Casier's (1960, fig. 3) generic separation of *C. carcharias* from the fossil species of *Carcharodon* (= *Procarcharodon*) was based on his view that it evolved from the variable *Oxyrhina* (= *Isurus*) during the upper Miocene. This hypothesis is not supported as *C. carcharias* is very closely related in morphology to the species of *Carcharodon*. Admittedly, there is a close resemblance in gross shape to justify Casier's claim that *C. carcharias* "arose through evolution as a variant of *O. hastalis* Agassiz" in the Miocene (Casier 1960, p. 13, fig. 3). If the presence of marginal denticulation (which

*O. hastalis* lacks) is to be regarded as a variable feature (which has appeared in some species of *Isurus*) then Casier's hypothesis that *C. carcharias* was derived from the *Isurus* lineage would be plausible. However, if marginal denticulation is a stable character, then it seems more logical to group this species under *Carcharodon*. Forms like *C. auriculatus* and *C. megalodon* with their long Tertiary history, show that denticulation is a constant morphological feature in the genus *Carcharodon*, not subject to variation through time. As the record for *C. carcharias* in New Zealand and Australia only begins in the Pliocene, this species apparently evolved elsewhere before the Pliocene, Chapman's (1918, p. 32) suggestion that it evolved earlier in North America but did not disperse to southern waters until the Pliocene seems logical, particularly as it (or its forerunners) has been recorded from the Eocene of America (Woodward 1889, p. 421) and the Miocene of Europe and America (Leriche 1936, table, p. 768). Thus *C. carcharias* probably had a long history as a distinct species, supporting the possibility that it was derived from *Carcharodon* stock during the Late Eocene or early Miocene rather than from *Isurus* stock during the Upper Miocene.

Casier (1960, fig. 3) suggested that *Procarcharodon* (= *Carcharodon*) arose in the Eocene as an offshoot from *Lamna* stock. However, the Eocene may have represented a period of evolutionary diversity with forms like *Otodus obliquus* Agassiz (*see* Eastman 1901, pl. 15, fig. 4) closely resembling *C. auriculatus* in gross morphology, and forms of *Isurus* also approaching *Carcharodon*, but it is not possible to relate any of these in an evolutionary sequence; all that can be said is that they show a convergent morphology. Although it could still be argued that *C. carcharias* evolved from *Isurus* in the Eocene, the case for treating *C. carcharias* as an offshoot from *Carcharodon* seems equally as probable. For this reason, in this paper *C. auriculatus* and *C. megalodon* have been retained in *Carcharodon* (s.lat.) and earlier nomenclature followed.

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# AUTHIGENIC PUMPELLYITE AND OTHER METAMORPHIC EFFECTS IN THE KYEBURN FORMATION, CENTRAL OTAGO

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ABSTRACT

Pumpellyite occurs as part of the metamorphic assemblage quartz-albite–pumpellyite–chlorite–sericite ± calcite ± ?stilpnomelane, in relatively shallow buried sandstones of the mid-Cretaceous Kyeburn Formation. In addition heulandite–clinoptilolite and montmorillonite occur in previously unrecorded tuff layers in the same formation, the former replacing both glassy matrix material and plagioclase feldspar. Coal rank and inferred depths of burial suggest that P–T conditions could have scarcely exceeded 50–130°C and 0·5–1 Kb.

## INTRODUCTION

The Kyeburn Formation of the Naseby District, Central Otago (Fig. 1), is a fanglomeratic sequence of middle to late Cretaceous breccia, conglomerate, and sandstone, with thin tuff beds, and is some 3600 m thick (Harrington 1955; Mutch 1963; Bishop and Force 1969). It overlies, and is derived from greywacke and semischist of the Torlesse Supergroup, from which it is separated by a profound metamorphic and tectonic unconformity. Apart from moderate tilting, the Kyeburn Formation is undeformed and is completely devoid of any penetrative structural features. Although relatively indurated in many places, some beds are sufficiently friable to be crumbled between the fingers, and the formation has been worked for gold by hydraulic sluicing in numerous places.

Tectonically equivalent late Mesozoic deposits of talus breccia, conglomerate and sandstone occur at several localities in the South Island, e.g., the Ohika Beds, Hawks Crag Breccia and Topfer Formation of South-west Nelson and Westland, and the Henley and Horse Range Breccias of Otago. Similarities are further enhanced by the association of minor acidic volcanics with the Ohika Beds (S. Nathan, pers. comm.), Topfer Formation (Suggate 1957, p. 24), the Horse Range Breccia (Steiner *et al.* 1959) and the Kyeburn Formation.

During a larger study of metamorphism of the underlying schist and greywacke (Bishop 1970), reconnaissance petrographic and X-ray examination of a few Kyeburn rocks revealed some important and unusual metamorphic or diagenetic effects. These may be divided into (a) development of quartz–albite–pumpellyite–chlorite–sericite ± calcite ± ?stilpnomelane assemblages in sandstones, and (b) montmorillonitic and zeolitic alteration in tuffs.

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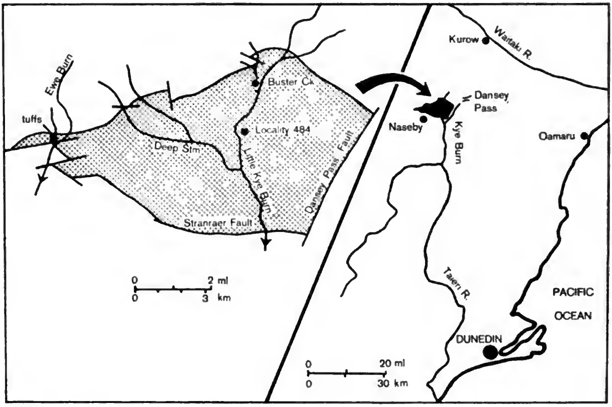


FIG. 1—Outcrop area and locality map of the Kyeburn Formation.

## ALTERATION IN SANDSTONES

A suite of hard, calcareous, carbonaceous, ?lacustrine sandstones and associated finer grained rocks were collected from the foot of a steep face in the Little Kye Burn at the Deep Creek confluence (field locality 484, S135/899795*, Fig. 2). The rocks consist of detrital quartz and albite, coalified plant fragments, minor epidote and sphene, and a few clasts of pelitic schist, and are set in a calcite cement. Authigenic phases additional to the carbonate cement are: fine grained quartz and albite, pumpellyite, chlorite, sericite, ?stilpnomelane, and sphene. Notes on the mineralogy of these phases follow:

### *Quartz and Albite*

Quartz and albite are the two most abundant clastic phases and positive identification of *clearly authigenic* grains of each is difficult. One or the other has been identified in most thin sections, however, and both are thought to be present in most rocks.

### *Pumpellyite*

Grains of blue-green pleochroic pumpellyite (Fig. 3) up to 0·1 mm in diameter constitute 3–5% of many rocks. Overgrowths of pumpellyite are

*Grid reference based on the sheet district of the 1: 63.360 topographical map series (NZMS 1) and the national thousand-yard grid shown on that series.

FIG. 2—Projecting bands of hard calcareous sandstone interbedded with fine sandstone and siltstone, probably representing a lacustrine phase of the Kyeburn Formation. Field locality 484, S135/899795, Little Kye Burn.

found on some grains of detrital epidote. Identification was confirmed by X-ray powder diffraction of a magnetic concentrate of OU 25418*. Pumpellyite was almost certainly present in some of the rocks from which the Kyeburn Formation was derived, and some has in fact been tentatively identified in clastic rock fragments (e.g., OU 25415). The textural characteristics and extensive recrystallisation of the Kyeburn rocks, however, strongly suggest that most of the pumpellyite, and the other phases described, are authigenic members of a stable metamorphic assemblage. Although typically associated with calcite, pumpellyite does not appear to be replacing it.

*OU numbers refer to the rock catalogue, Geology Department, University of Otago.

*Chlorite*

Strongly pleochroic, optically negative chlorite ($\alpha$ = pale yellow-green, $\beta = \gamma$ = deep bluish-green) with relatively high birefringence occurs in most rocks as tiny flakes or plates averaging 0·05 mm in diameter. The basal spacing, determined from the 004 reflection of chlorite from a thin layer of intercalated lenses of authigenic chlorite-sericite-calcite and carbonaceous material (OU 25416) is 14·12Å. The optical data indicate an Fe-rich chlorite (Albee 1962).

*Sericite*

Sericite is a ubiquitous component of the sandstones. Relatively strong 114 and $11\bar{4}$ reflections on powder diffraction runs of material from OU 25416 (see above) indicate the sericite is a 2M polymorph; the basal spacing of 19·86Å may indicate a phengitic composition.

*Sphene*

Minute granules of sphene are widely distributed.

*Carbonates*

Fine grained calcite is abundant in the rocks from this locality. It occurs as a cementing medium in sandstones, as almost monomineralic thin dark-coloured layers (calcite, minor carbonaceous material and quartz) parallel to

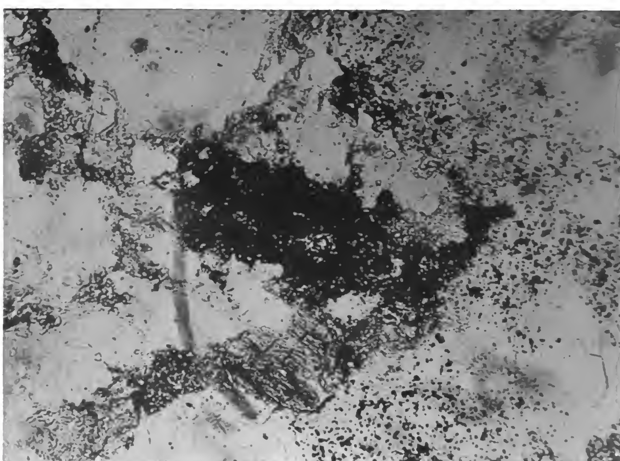

FIG. 3—Authigenic grain of dark, spongy pumpellyite (centre) in calcareous sandstone, OU 25425 (× 120, plain polarised light).

bedding, or as isolated crystals in coal. Unusually coarse calcite is often associated with plant fragments (e.g., OU 25413). Calcite cement may constitute up to 40% of the sandstones. A small patch of unidentified pale pinkish-brown pleochroic ($\omega > \epsilon$) carbonate occurs in OU 25414. Dolomite was not detected by X-ray diffraction, selective staining, or refractive index methods.

*Stilpnomelane*

Radiating sprays of a pleochroic (pale yellow to dull deep green or pale yellow to dark brown) mineral, sometimes showing a weak parting perpendicular to the basal cleavage, are tentatively referred to stilpnomelane or possibly oxy-chlorite (OU 25415, 25414).

## ALTERATION IN TUFFS

Brownish-grey, 5–75 cm thick, tuff layers containing conspicuous biotite and with a distinctive waxy appearance in hand specimen, occur interbedded with conglomerates in the Ewe Burn (S135/811793). In thin section (OU 25411) the tuffs consist of ragged flakes of biotite up to 1 mm in diameter (25%), smaller (0·2 mm), rounded or angular grains of quartz (15%) and oligoclase (5%), and wispy muscovite (2%). Rather rounded, corroded grains of iron ore, rare zircon, and a few fragments of fine grained pelitic schist are also present. The muscovite and some of the rock fragments may be of sedimentary rather than primary or xenolithic origin, as rounded pebbles occur in the lower part of the thickest tuff band. The reddish-brown groundmass (50%) contains a few relict shard structures and is largely altered to fine grained montmorillonite.

Float boulders of another biotite-bearing tuff were collected nearby. In hand specimen they are pale cream or white rocks, speckled with biotite, and are sometimes distinctly graded. The waxy lustre of the montmorillonitic rocks is lacking and instead these tuffs have a dull or earthy appearance. OU 25410 contains the same primary phases as OU 25411, although in different proportions (calcic oligoclase 30%, quartz [including a few grains with the bipyramidal terminations of originally $\beta$ quartz] 25%, biotite 5%). Iron ore, schist xenoliths, green hornblende and zircon make up a further 3%. The groundmass, about 40% of the rock, has a vitroclastic texture and is largely altered to a weakly birefringent mineral of the heulandite–clinoptilolite series. Zeolitic replacement of feldspar has also occurred (Fig. 4); this phase shows negative optical elongation and appears to have slightly higher refractive indices than the length-slow zeolite of the groundmass. Both properties are consistent with a more heulandite-rich variety in the feldspars than in the groundmass. Replacement of plagioclase feldspar by zeolites of the heulandite–clinoptilolite series has been reported by Hay (1963), but appears to be rare.

## COAL RANK

Proximate analyses of three coal samples from the Kyeburn Formation were made by the Coal Research Association and the rank number (Suggate 1959) determined. The coal ranks (converted to SI units) are plotted against

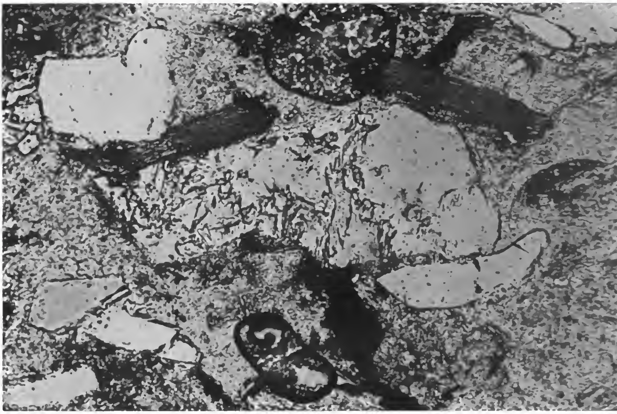

FIG. 4—Zeolitic replacement of feldspar in biotite tuff from the Ewe Burn. (OU 25410, × 120, plain polarised light).

inferred depth of burial in Fig. 5, on which the ideal depth–rank relationship and the position of locality 484 are also indicated. Two samples, one above and one below locality 484, have ranks consistent with the inferred depth of burial. The rank of the third is considerably less than expected, and may be due to a weathered sample, or to its position closer to the distal feather-edge of the Kyeburn Formation.

## DISCUSSION

Depth of burial of the tuffs from the Ewe Burn is difficult to estimate, as the sequence is severely faulted in this area; about 2000 m appears probable. Zeolitic or montmorillonitic alteration of volcanic glass is widespread in such environments (e.g., Hay 1966; Coombs 1971).

The calcareous rocks of the Little Kye Burn are about 2500 m below the top of the section exposed, and an additional 1500 m may have been faulted out. Even allowing an overburden of 4000 m (probably excessive in view of the coal ranks), temperatures at this locality are unlikely to have exceeded 130°C. An unusual chemical environment during and after deposition is indicated by the restriction of calcareous rocks to the immediate area of locality 484. A pumpellyite–chlorite–sericite assemblage, however, also occurs in OU 25417, a friable blue-grey, gritty, carbonate-free, mildly carbonaceous sandstone from Buster Creek* (S126/908824) within a few hundred metres

*Not named on Sheet S126, but shown by Williamson (1939).

of the top of the formation, indicating that pumpellyite-bearing assemblages may be widespread in the Kyeburn Formation, and are not restricted to the calcareous facies.

Pumpellyite is a common phase in the deeper buried parts of low-grade metamorphic sequences, where it is usually overlain by at least 3–5 km of pumpellyite-free rocks (e.g., Coombs 1954; Packham and Crook 1960; Otalora 1964). It has been reported from shallow depths (50–600 m) and low *present day* temperatures (100–230°C) in drillholes in geothermal areas in Iceland (Sigvaldasson 1963) and has also been reported as a low temperature, low pressure deuteric mineral in cavities in the Prospect Intrusion, New South Wales (Raam *et al.* 1969). Pumpellyite also appears to be a relatively common alteration product of biotite in granitic and dioritic rocks (Struwe 1958).

The pumpellyite occurrences in the Kyeburn Formation appear to represent conditions of low pressure ($P_{total} = c.\ 0{\cdot}5$–1Kb) and low temperature (50–100°C), in rocks containing abundant free carbon. In addition the presence of detrital pumpellyite may have catalysed the early nucleation of authigenic pumpellyite, although no evidence of overgrowths was detected optically. Estimation of other chemical conditions prevailing during metamorphism must await the results of more detailed study of these rocks.

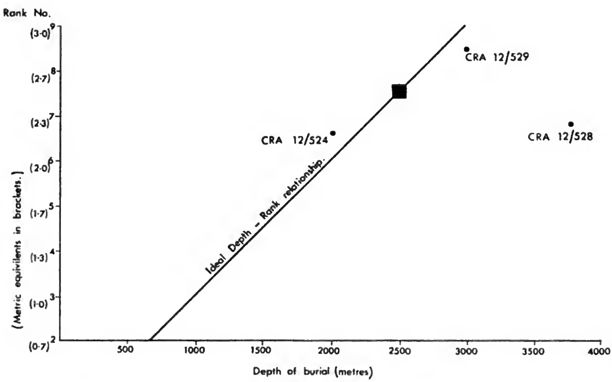


FIG. 5—Relationship of coal rank to inferred depth of burial. Stratigraphic position of locality 484 indicated by square, position of coal samples by dots. CRA 12/524 from S136/900783, Little Kye Burn, 220 m above Deep Creek confluence; CRA 12/528 from S126/916832, Little Kye Burn, 550 m upstream from ford; CRA 12/529 from S126/906812, Little Kye Burn, 1400 m downstream from ford.

ACKNOWLEDGMENTS

The writer is indebted to Professor D. S. Coombs, Drs C. A. Landis and W. A. Watters, and J. R. Boles, for discussion and assistance during the preparation of this manuscript, and also to the Coal Research Association for the Kyeburn coal analyses.

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# ROTOEHU ASH AND THE ROTOITI BRECCIA FORMATION, TAUPO VOLCANIC ZONE, NEW ZEALAND

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(*Received 18 May 1971, revised 31 August 1971*)

## ABSTRACT

The widespread shower-bedded deposits of Rotoehu Ash consist of multiple airfall pyroclastic units (tephras) which underlie, are interbedded within and mantle the less widespread rhyolitic pyroclastic flow deposits of the Rotoiti Breccia Formation (late Pleistocene). No significant time intervals are recorded within the Rotoehu Ash, showing that the thick pyroclastic flow breccia deposits enclosed by the Rotoehu Ash units were also erupted in a short time interval.

Recognition of the multiple nature of Rotoehu Ash has proved necessary in correlation of breccia deposits in the Rotorua area.

## INTRODUCTION

In their recent study of late Pleistocene volcanic ash deposits in the central North Island, Vucetich and Pullar (1969, pp. 805–6) named a major tephra deposit as Rotoehu Ash. They considered Rotoehu Ash was the widespread, "prominently shower-bedded ash associated with the terminating eruptions of Rotoiti Breccia", and that Rotoehu Ash rested "without significant time interval on recognised deposits of Rotoiti Breccia".

Rotoiti Breccia was first named in Sheet 5, "Geological Map of New Zealand 1:250,000" (Healy *et al.* 1964). It was later briefly described by Ewart and Healy (1965, p. 23): "A series of pumice breccias laid down as hot avalanche deposits form an extensive fan covering an area of approximately 325 square miles north of Lakes Rotoiti, Rotoehu and Rotoma" Thompson (1968, p. 1190) described the Rotoiti Breccia as including all the airfall ash, breccias, and tuffaceous siltstone overlying the weathered surface of the Mamaku Ignimbrite and beneath the Rotorua Subgroup of Holocene volcanic ash beds. However, Vucetich and Pullar (1969) recognised nine significant time intervals (buried soils) between the Rotorua Subgroup and the deeply weathered brown ashes which overlie the Mamaku Ignimbrite. They identified the lowest of these buried soils as marking the top of the Rotoiti Breccia Formation and renamed the immediately overlying sequence of block and lapilli beds as the Mangaoni Lapilli Formation. These interpretations of the Rotoiti Breccia Formation are shown in Fig. 1, together with that presented in this paper. The writer follows Vucetich and Pullar (1969) in regarding the top of the Rotoiti Breccia Formation as the buried soil underlying the Mangaoni Lapilli Formation. The breccia deposits beneath this upper contact consist of rhyolitic pyroclastic flow units of unwelded ash, lapilli, and blocks, plus shower-bedded airfall units (tephras).

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| Thompson (1968) General sequence, Lakes Rotorua - Rotoma | | Vucetich and Pullar (1969) | | This paper | |
|---|---|---|---|---|---|
| Lithology | Formation | Lithology | Formation | Lithology | Formation |
| Brown volcanic ash | ROTORUA SUB-GROUP | | ROTORUA SUB-GROUP | | ROTORUA SUBGROUP |
| Loosely compacted pumice breccia and inter-stratified airfall ash beds | ROTOITI BRECCIA | | OKAREKA ASH<br>TE RERE ASH<br>ORUANUI ASH | | OKAREKA ASH<br>TE RERE ASH<br>ORUANUI ASH |
| | | | MANGAONI LAPILLI FORMATION (5 members) | | MANGAONI LAPILLI FORMATION (5 members) |
| | | Shower-bedded ash | Rotoehu Ash | Cross-bedded ash<br>Shower-bedded ash | Rotoehu Ash<br>ROTOITI BRECCIA FORMATION |
| Tuffaceous siltstone | | | ROTOITI BRECCIA FORMATION | Flow units<br>Shower-bedded ash<br>Flow units<br>Shower-bedded ash | |
| Brown clay weathered surface | | Brown ashes | | Brown ashes | |
| Ignimbrite | MAMAKU IGNIMBRITE | | | Ignimbrite | MAMAKU IGNIMBRITE |

FIG. 1—A comparison of interpretations of Rotoiti Breccia Formation in the pyroclastic column

The term "pyroclastic flow" is used to include all those types of eruption variously referred to as "nuées ardentes", "glowing avalanches", "ash flows", etc. "Tephra" is used to denote airfall deposited pyroclastic material.

In Vucetich and Pullar (1969), shower-bedded Rotoehu Ash tephra was said to overlie the pyroclastic flow deposits of Rotoiti Breccia. However, Smith (1963, p. 110) states: "Almost all ash flow successions are preceded by airfall eruptions". This eruptive sequence has been observed in the Aira and Ata pyroclastic flow deposits in Japan (Aramaki and Ui 1966), the Krakatau eruptions of 1883 (Williams 1941), and the Crater Lake, Oregon, deposits (Williams 1942). New Zealand examples include the Oruanui Breccia overlying shower-bedded Oruanui Ash (Vucetich and Pullar 1969), and pyroclastic flow deposits of the Upper Taupo Pumice Members overlying the Taupo Lapilli and other airfall members of the Taupo Pumice Formation (Healy 1964). New evidence is here presented for a similar eruptive sequence within the Rotoiti Breccia Formation. Shower-bedded tephras have recently been found beneath, and intercalated between Rotoiti Breccia flow units, and are correlated with the widespread Rotoehu Ash.

## STRATIGRAPHY

Geographic names and locations of sections are shown in Fig. 2.

Vucetich and Pullar (1969, p. 805) described Rotoehu Ash from a reference section on the Whakatane–Ohope Road (N69/450243*, *see* Fig. 2), where the lower beds of the Rotoehu Ash consist of distinctively grey coloured, finely shower-bedded coarse and fine ashes, which overlie brown weathered ash. The fine shower-bedding of the basal Rotoehu Ash at the Ohope reference section is in strong contrast with the weak shower-bedding of the tephra which conformably mantles the upper Rotoiti Breccia flow deposits in sections studied by the writer at Paengaroa (N67/843353), Roydon Downs (N67/885358), and Otamarakau (N68/078401). The mantling tephra at these upper sections consists of 1–2 m of weakly shower-bedded coarse ash and lapilli, and was also found to contain biotite. Biotite has not previously been recognised in either Rotoiti Breccia (cf. Ewart 1968, p. 529) or Rotoehu Ash, and appears to be restricted to the tephra unit overlying the Rotoiti Breccia flow deposits. This tephra unit is here informally named the "mantling tephra".

[At the Paengaroa and Roydon Downs sections the shower-bedded mantling tephra is conformably overlain by finely stratified and cross-bedded ashes, comprising member d of Healy *et al.* (1964). The origin of these cross-bedded deposits is not yet certain (base surge?, water deposited?) but their distribution appears largely restricted to that of the Rotoiti Breccia flow deposits.]

During recent investigation of the Rotoiti Breccia, sections were found exposing the basal contact of the Formation on deeply weathered brown ashes overlying Mamaku Ignimbrite. The reference section is at Saunder's Track (N67/818347, Fig. 3), where 30+ m of unstratified Rotoiti Breccia flow

*Grid references are based on the national 1,000 yard grid of the 1: 63,360 topographical map series (NZMS 1).

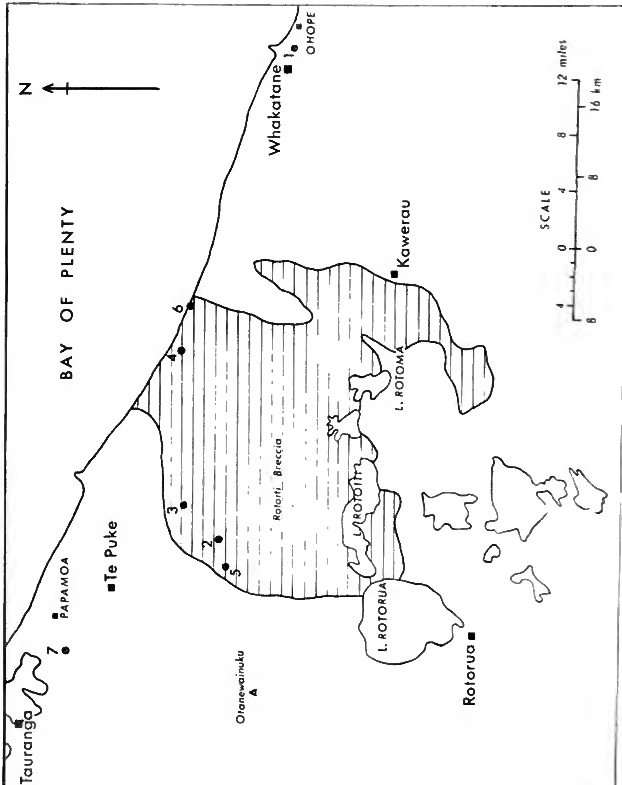


FIG. 2—Generalised distribution of Rotoiti Breccia flow deposits (modified from Healy *et al.* 1964), and locations of tephra sections (denoted by numbers) referred to in text. 1–Rotoehu Ash reference section, Ohope (Vucetich and Pullar 1969); 2–Paengaroa; 3–Roydon Downs; 4–Otamarakau; 5–Saunder's Track; 6–Mimiha Stream; 7–Reids Road.


FIG. 3—Saunder's Track reference section. Rotoiti Breccia flow unit (rb) overlying shower-bedded basal tephra. Base of spade rests at paleosol underlying the basal tephra; spade is 1 m in length. (wa–white fine ash bed; ga–grey ash bed).

deposits overlie without time break 2 m of distinctively coloured and finely shower-bedded ash, here informally named "basal tephra". The basal tephra rests on a highly carbonaceous paleosol (buried soil). The section is as follows:

| | |
|---|---|
| Rotoiti Breccia (flow deposits) | >30 m (basal 7 m exposed) unstratified grey ash and pumice lapilli, containing some charred wood. |
| | ——— sharp contact ——— |
| Shower-bedded ash and lapilli (basal tephra) | 38 cm multiple shower-beds of grey, white, brown, and black fine ashes.<br>71 cm yellow-brown coarse ash containing small lapilli.<br>46 cm white very fine ash, containing pisolites.<br>15 cm grey medium ash. |
| | ——— sharp contact ——— |
| brown ashes Mamaku Ignimbrite | Highly carbonaceous paleosol containing much charred wood, passing downwards into strongly weathered brown ash beds over ignimbrite. |

This sequence demonstrates that the first products of the Rotoiti Breccia eruptions to reach the Saunder's Track locality were airfall pyroclastics which fell on to strongly developed soils and abundant vegetation. The contact between the basal tephra and overlying breccia flow deposits is without weathering break and generally conformable, although in places the immediately overlying flow has slightly eroded the uppermost tephra bed (Fig. 3). The pisolitic white fine ash bed is a probable correlative of the basal tuffaceous siltstone of the Rotoiti Breccia, as described by Thomson (1968, p. 1190) and dated at >41,000 years B.P., 67% probability (NZ $^{14}$C643).

The basal tephra has also been identified at Mimiha Stream (N68/149366) in the Matata to Otamarakau coastal section, described by Healy and Ewart (1965, p. 132). The basal tephra rests on a paleosol developed on pink silts and sands, and is conformably overlain by unstratified Rotoiti Breccia flow deposits.

Prominent beds within the basal tephra at the Saunder's Track section are the basal grey ash and immediately overlying pisolitic white fine ash bed. Both these distinctive beds are readily correlated with basal beds of the Rotoehu Ash at sections beyond the margins of the Rotoiti Breccia flow deposits. Rotoehu Ash is well exposed at Reids Road, Papamoa (N58/742546, Fig. 4),

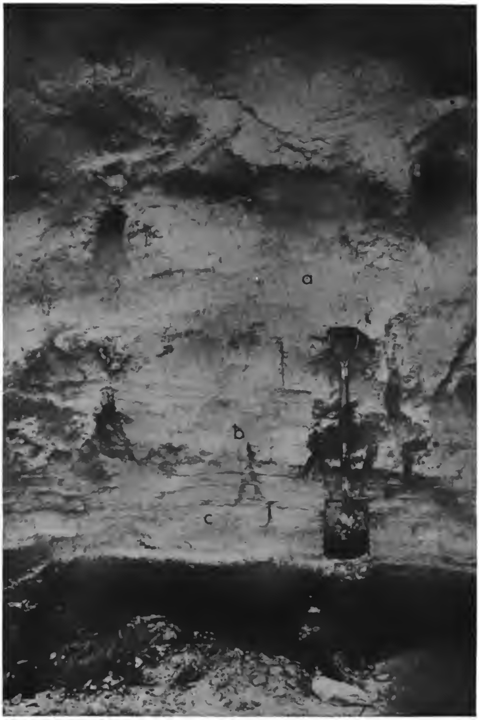


FIG. 4—Rotoehu Ash sequence at Reids Road, Papamoa. Base of spade at lower contact of Rotoehu Ash (a–upper beds; b–intercalated beds; c–basal beds).

some 16 km north-west of the north-west margin of the Rotoiti Breccia flow deposits. The section is as follows:

| | | |
|---|---|---|
| | | Mangaoni Lapilli Formation |
| Rotoehu Ash | | ——————irregular upper contact—— |
| | (upper beds) | 66 cm weakly shower-bedded coarse ash and lapilli, containing sparse biotite. |
| | | - - - - - - - - - - - inferred break - - - - |
| | (intercalated beds) | 18 cm white and light brown fine ash.<br>8 cm weakly shower-bedded medium to coarse ash.<br>23 cm white and light brown fine ash. |
| | | - - - - - - - - - - - inferred break - - - - - - |
| | (basal beds) | 15 cm finely shower-bedded, multiple grey fine and medium ashes.<br>15 cm shower-bedded, grey and white medium to coarse ash.<br>6 cm white fine ash.<br>2·5 cm grey medium ash. |
| | | —————— sharp contact —————— |
| | | weathered sands |

At this section the upper, biotite-bearing ash beds are correlated with the biotite-bearing tephra overlying Rotoiti Breccia flow deposits in the Roydon Downs and other sections where mantling tephra is exposed. The distinctively shower-bedded basal ash beds are readily correlated with the basal tephra underlying the Rotoiti Breccia flow deposits at the Saunder's Track section, the grey ash and white fine ash beds being particularly apparent. Intercalated between the distinctive upper and basal beds are further non-distinctive shower-beds which are collectively correlatives of tephras interbedded between pyroclastic flow units of the breccia, and are here informally named "intercalated beds". No weathering or erosional breaks are apparent within the Rotoehu Ash sequence at the Reids Road section, or within any other exposure of the Rotoehu Ash (Vucetich and Pullar pers. comm.).

From the sections described above, the Rotoehu Ash is seen to consist of multiple tephra units within the Rotoiti Breccia Formation. The tephra units underlie, are interbedded with, and mantle the pyroclastic flow units of the Rotoiti Breccia. The distinctive basal shower-bedded ash—a criterion for recognition of the Rotoehu Ash (Vucetich and Pullar 1969, p. 806)—is the widespread correlative of the basal tephra underlying the Rotoiti Breccia flow deposits. An idealised diagram of the relationship between the Rotoiti Breccia Formation and Rotoehu Ash is shown in Fig. 5 .

An approximate calculation of the erupted volume of the Rotoehu Ash deposits, using isopachs generalised from Vucetich and Pullar (1969), gave a volume $\geqslant 50\ km^3$. An equally approximate calculation of the volume of the Rotoiti Breccia flow deposits also gave $\geqslant 50\ km^3$, indicating that about half of the Rotoiti Breccia Formation was erupted as widespread airfall tephra.

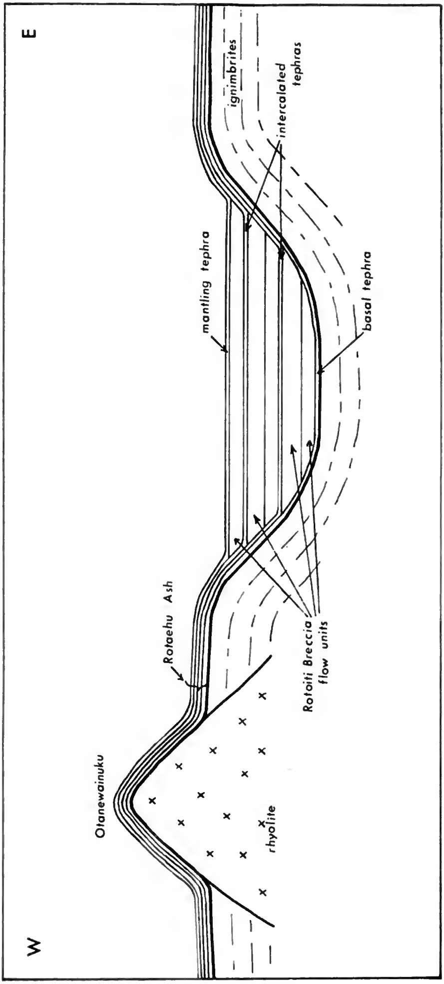


FIG. 5—Diagram of an idealised sequence showing relationship of Rotoiti Breccia flow deposits and Rotoehu Ash. In this northern sector, the flow deposits of the Rotoiti Breccia appear to occupy a sag in the ignimbrite terrain. (Rotoehu Ash thickness and vertical scale much exaggerated).

Recognition of the multiple nature of Rotoehu Ash has proved necessary in correlation of local pyroclastic flow deposits. Vucetich and Pullar (1969, p. 803) described pumice ash breccias on the eastern shore of Lake Rotoma and at Kawerau as overlying Rotoehu Ash and therefore postdating Rotoiti Breccia. However, the writer considers the tephra exposed at these localities to be the basal tephra of the Rotoiti Breccia Formation, and that the overlying pyroclastic flow deposits are in fact Rotoiti Breccia.

Isopachs of Rotoehu Ash thickness (Vucetich and Pullar 1969, fig. 14) will be in error on the Rotoiti Breccia flow deposits as only the mantling tephra thickness will have been measured. True Rotoehu Ash isopachs, and interpretation for eruption source, would require exposure of the entire depth of the Rotoiti Breccia Formation and summation of the tephra thicknesses.

Recent work (Nairn 1971) has shown that eruption of the Rotoiti Breccia was followed, without significant time interval, by the smaller Earthquake Flat Breccia pyroclastic eruptions. Widespread tephra units associated with the Earthquake Flat Breccia have been identified in several Rotoehu Ash sections in the Rotorua–Taupo area. .The upper Rotoehu Ash in other areas may also include components from the Earthquake Flat eruptions.

## INTERPRETATION

The Rotoehu Ash deposit at the Reids Road section is entirely conformable and shows no weathering intervals between the basal, intercalated, and upper beds. Deposition of the beds must have occupied only a short time interval with no pauses long enough for weathering or erosion to occur. Correlation of these tephra beds with the tephras underlying, interbedded with, and mantling the thick Rotoiti Breccia pyroclastic flow deposits shows that the eruption of the enclosed flow units occupied an equally short period of time. The correlations indicate that the entire Rotoiti Breccia Formation was erupted in a short time interval and therefore probably constitutes a single cooling unit.

The Rotoiti Breccia Formation consists of numerous flow units and interbedded tephra units, representing different modes of deposition from the same eruptive centre.. Distinctive lithologies and mineralogies are present in the tephra units, and further detailed study will enable the naming of tephra and flow unit members.

## ACKNOWLEDGMENTS

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# LATE PALEOZOIC GLACIAL VALLEY AT ALLIGATOR PEAK, SOUTHERN VICTORIA LAND, ANTARCTICA

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ABSTRACT

A north-east trending Paleozoic glacial valley has been cut in Devonian Aztec Siltstone at Alligator Peak, southern Victoria Land. The valley, which is exposed in a north-west–south-east cross-section, has walls as steep as 50°, and is filled to a depth of 80 m with glacial and fluvial sediments of the Metschel Tillite (Permian-Carboniferous (?)).

The oldest fill is complexly folded and comprises locally brecciated glacial and fluvioglacial sandstone, conglomerate, and claystone. This is overlain by fluvial sandstone, and the fill is completed by a lacustrine siltstone lens. Strikes of fold axes and other linear features in the valley trend north-east, subparallel to the strike of the valley wall, indicating that the oldest fill was deposited by slumping of the valley wall and not by "ice shove" along the length of the valley.

## INTRODUCTION

A small Late Paleozoic valley filled with glacial and fluvial sediment was found at Alligator Peak, southern Victoria Land (Fig. 1), by members of the 1970–71 Victoria University of Wellington Antarctic Expedition (Barrett *et al.* 1971). The valley is cut in flat-lying Devonian Aztec Siltstone (Beacon Supergroup, Table 1), and is overlain by Permian Weller Coal Measures. The

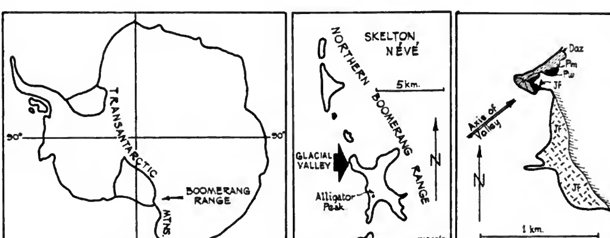


FIG. 1—Alligator Peak, Boomerang Range, south Victoria Land. Arrow showing axis of Late Paleozoic valley also indicates viewing direction of Fig. 2 to 4. Daz = Aztec Siltstone; Pm = Metschel Tillite; Pw = Weller Coal Measures; Jf = Ferrar Dolerite.

TABLE 1—Stratigraphy of the Beacon Supergroup in Southern Victoria Land (*after* McElroy 1969; McKelvey *et al.* 1970)

| | | | |
|---|---|---|---|
| BEACON SUPERGROUP | VICTORIA GROUP | Triassic | Lashly Formation |
| | | Permian or Triassic | Feather Conglomerate |
| | | Permian | Weller Coal Measures |
| | | | Pyramid Erosion Surface |
| | | Permian - ?Carboniferous | Metschel Tillite |
| | | | Maya Erosion Surface |
| | TAYLOR GROUP | Middle to Upper Devonian | Aztec Siltstone |
| | | | Beacon Heights Orthoquartzite |
| | | | Arena Sandstone |
| | | Devonian and possibly older | Altar Mountain Formation |
| | | | Heimdall Erosion Surface |
| | | | New Mountain Sandstone |
| | | | Kukri Surface |
| | | Lower Paleozoic-Precambrian | BASEMENT COMPLEX |

valley fill is part of the widespread but discontinuous Metschel Tillite (McKelvey *et al.* 1970) in the Boomerang Range–Skelton Neve area, and is equivalent to the thicker and more closely studied glacial deposits at the same stratigraphic position farther south in the Transantarctic Mountains (Long 1965; Frakes *et al.* 1966; Schmidt and Williams 1969; Lindsay 1970; Barrett *et al.* in press).

The Metschel Tillite of southern Victoria Land normally is no more than 40 m thick and comprises one or two thick massive beds of poorly sorted fine sandstone with scattered pebbles and boulders as much as a metre across. However the formation locally includes well-bedded and moderately-sorted sandstone and siltstone. The glacial origin of the massive beds is indicated by the poor sorting, and the variety of clast lithologies (McKelvey *et al.* 1970, in press; Pinet *et al.* 1971) and by the finely grooved and striated surface on the Aztec Siltstone found this season at the base of the formation at Mount Metschel.

## DESCRIPTION

The glacial valley at Alligator Peak (Fig. 1) trends north-east (Table 2) and is exposed in a north-west–south-east cross-section. The north-western wall (Fig. 2) is the only part exposed, and dips at about 50°; the contact between the valley fill and the Aztec Siltstone is knife-sharp.

The oldest valley fill is sandstone, conglomerate and claystone of glacial and fluvioglacial origin. This is overlain by a massive sandstone of probable fluvial origin and the fill is completed by a lacustrine siltstone lens.

The glacial and fluvioglacial fill includes poorly bedded or unbedded massive tillite, claystone, and well-bedded feldspathic sandstone and conglomerate. The clasts in the conglomerate, which are as much as 40 cm across, are rounded and mainly granitic, like those in the tillite. The relationships between the several lithologies have been complicated by penecontemporaneous slumping (Fig. 2, 3). The slumping has not affected the layer of thinly bedded coarse sandstone adjacent to the valley wall (Fig. 3), or the 20 cm thick dense black finely vesicular bed stratigraphically above it. Sharp upper and lower contacts suggest that the black rock was a shallow intrusion,


FIG. 2—Cliff-section through exposed northern side of Late Paleozoic glacial valley, looking east. Arrow indicates figure standing on valley wall for scale. Corners indicate area of Fig. 3.

TABLE 2—Directional Features Measured in Late Paleozoic Glacial Valley and Sediments

| Locality (*see* Fig. 3) | Mean | Number | Description |
|---|---|---|---|
| a | 20° | 3 | Strike on valley wall |
| b | 37° | 1 | Strike of bedding |
| c | 27° | 1 | Strike of bedding |
| d | 22° | 1 | Strike of thrust surface |
| e | 48° | 6 | Lineations (?grooves) on sandstone beds |
| f | 20° | 5 | Lineations (shear lamellae) |
| g | 48° | 2 | Flow fold axes |
| h | 44° | 6 | Planar cross-bedding in massive sandstone |

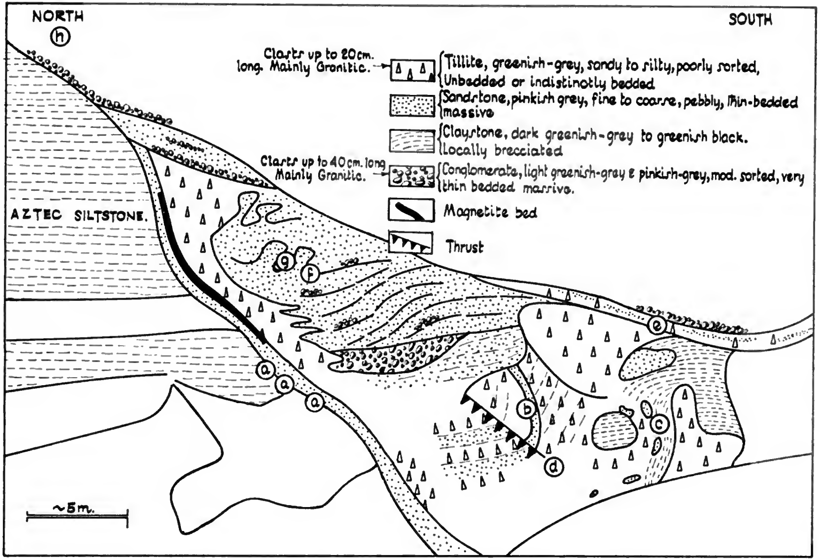


FIG. 3—Sketch of cliff section through glacial valley, taken from Fig. 2 and field notes, showing the relationship between the various lithologies in the lower part of the valley. Directions in Table 2 were measured at localities a–g.

but it is composed almost entirely of magnetite (B.P. Kohn, pers. comm., Sept. 1971).

The overlying massive fine sandstone (Fig. 2, 4), which is 40 m thick, forms a shallow syncline over the valley. The northwest limb strikes parallel to and rises above the valley's north-west wall where the dip increases to 30°. The lower contact is erosional near the valley wall but is interfingering toward the centre of the valley (Fig. 2). The unit is parallel-bedded except for a pocket of planar cross-bedding near the base.

The massive sandstone is overlain by a lens of shaly, laminated, greenish grey siltstone as much as 7 m thick (Fig. 4). Laminae of medium grained white quartzose sand with well developed lobate sole marks are common in the lower and middle parts of the lens, and may have been deposited by turbidity currents. Convolute bedding, slump folds on a scale of 1 to 3 cm and ripple lamination were also found. Scattered granite and quartz pebbles occur in a silt matrix at the northern end of the lens in the middle of the unit. Laminae at the base of the lens follow the shape of the syncline but those at the top are horizontal.

The siltstone is disconformably overlain by horizontal beds of medium- to coarse-grained trough cross-bedded sandstone with lenses of granite and quartz pebbles, and boulders, typical of the lower part of the Weller Coal Measures in the area.

## DISCUSSION

The steepness of the wall of the glacial valley and the sharpness of the contact suggest that the valley was carved by glacier ice. The absence of any sign of creep in the siltstone of the valley wall indicates that the valley was filled with sediment as the ice front retreated or shortly after, because the Aztec Siltstone was not much lithified at the time of glaciation. (At Mt Metschel, 20 km north, the siltstone immediately beneath the tillite has been deformed into isoclinal folds (McKelvey *et al.* in press, fig. 5), presumably as a result of ice movement.)

The severely folded tillite and sandstone that form the lowest exposures in the valley were deposited by slumping off the valley wall, not by "ice shove" along the length of the valley. The strikes of bedding, fold axes and other directions in the valley fill are subparallel to the strike of the valley wall (Table 2, Fig. 3 a–g), indicating movement normal to the trend of the valley. The interfingering relationship with sandstone and adjacent tillite indicates local contemporaneous deposition from streams and ablating ice. The large feldspathic sandstone and conglomerate body near the valley wall was probably deposited from streams flowing along the side of the valley as the ice front retreated. Local northward dips in the sandstone may have resulted from subsidence after melting of ice trapped beneath the sand. This may have also caused the local flow folding with axes parallel to the adjacent valley wall (Fig. 3, g; Table 2).

Slump-folded stratified beds in the Metschel Tillite have also been seen at Kennar Valley, 80 km north-north-east of Alligator Peak (Dr B. C. McKelvey, University of New England, pers. comm., Feb., 1971).

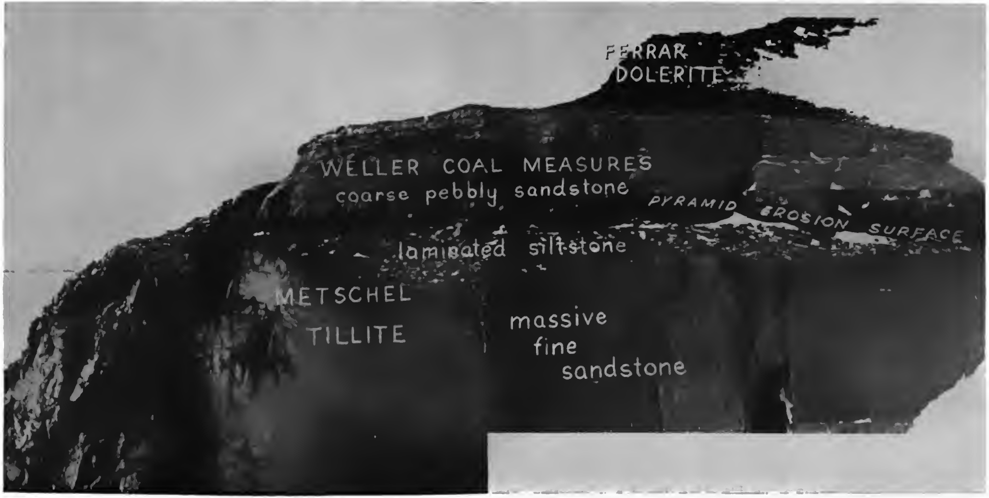


FIG. 4—Cliff-section above glacial valley shown in Fig. 2 and 3, looking east. Sandstone and siltstone of the Metschell Tillite Formation are disconformably overlain by medium to coarse-grained pebbly sandstone of the Weller Coal Measures. The upper sandstone bluff is 19 m high.

The massive sandstone overlies the older valley fill with local conformity. This, and the synclinal form of the body, indicate that it was deposited very shortly after the earlier fill was emplaced, and before there had been much compaction, possibly as the ice was still retreating. A set of cross-bedding measurements 3 m above the base of the sandstone indicate north-easterly current flow (Table 2, h). If the ice front had not left the valley this also represents the direction of the ice flow. Most compaction appears to have taken place during the deposition of the overlying siltstone lens (Fig. 4), whose oldest laminae are curved but whose youngest laminae are still horizontal. The overlying Weller Coal Measures is similar in its coarse-grained and fluvial character and in its erosional lower contact to occurrences elsewhere in the area, where it overlies thin Metschel Tillite or the underlying Aztec Siltstone. By Weller time the influence of the glacial valley on local sedimentation had clearly ended.

## ACKNOWLEDGMENTS

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I am also grateful to Messrs B. P. Kohn and R. H. Grapes, Department of Geology, Victoria University, and Dr P. N. Webb, N.Z. Geological Survey, for reviewing the manuscript.

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# BONE OF A PRESUMED ODONTOPTERYGIAN BIRD FROM THE MIOCENE OF NEW ZEALAND

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(*Received 8 February 1971, revised 20 January 1972*)

ABSTRACT

A partial right humerus of a bird of probable Waiauan (middle to late Miocene) age, from the coastal cliffs of North Canterbury, and tentatively considered to belong to the order Pelicaniformes: sub-order Odontopterygia (bony-toothed birds) is described.

## INTRODUCTION

The bird bone discussed below was found by Kevin Tyree in January 1969, in a fine-grained sandstone. The location is about 6 m above mean sea-level in coastal cliffs about 3 km (2 miles) north-east of the Waipara River mouth, North Canterbury. (Sheet S68 (3rd ed.) of the NZMS 1 series, grid reference 177041).

Kevin Tyree later presented the bone to the Canterbury Museum.

A further visit to the locality was made on 11 October 1969 by the Tyree family, Dr S. P. Welles, of the University of California, Mr D. R. Gregg, Geologist at the Canterbury Museum, and myself, but no more bone was found. On this visit Mr Gregg took a sample of matrix, as close as possible to where the bone had been embedded, and this was examined for foraminiferal content by Dr N. de B. Hornibrook, N.Z. Geological Survey, Lower Hutt. Results are given below. The coastal cliff sediments were mapped by Wilson (1963) as Southland to Taranaki series, and by Gregg (1964) as Southland series.

## AGE

The age of the sediments is "very probably" Waiauan (Southland Series, middle to late Miocene according to Dr N. de B. Hornibrook. His conclusion is based on an analysis of the matrix sample, in which he determined few in number of the following shallow-water Foraminifera: *Florilus parri* (Cushman), Waiauan–Recent; *Rectobolivina striatula* Cushman, Waiauan–Recent; *Loxostromum pakaurangiense* Hornibrook, Otaian–Waiauan; *Elphidum* sp.; and *Notorotalia* cf. *spinosa* (Chapman).

Canterbury Museum Catalogue Number AV24,960: partial right humerus. Locality given above. The matrix is a fine-grained sandstone correlated with the Double Corner Shellbeds of Gregg (1959). The bone consists of the

distal end and part shaft of a right humerus of a very large bird. It is of a light-brown colour. When found, it was fractured into five segments, now glued together.

The nomenclature used in describing the bone is based mainly on Fürbringer (1888) and diagrams (Fig. 1) are included for the convenience of readers unfamiliar with bird osteology.

Portions of the distal end and of a projection or "flange" on the upper or lateral edge are abraded. The bone has undergone a little crushing, and possible slight distortion, presumably through pressure. The total length of the surviving humerus is 45·5 cm.

The shaft proper varies in depth (outer, or palmar side to inner, or anconal side) from 2·2 to 2·4 cm. By "outer side" is meant the side (*not* edge) of the bone which faces away from the body, and by "inner", the side of the bone which is nearest to the body. .

The transverse width of the shaft varies from 2·6 cm at a point 38·7 cm from the distal end (below the fossil barnacle visible in Fig. 2C), to 3·5 cm at the beginning of the flange. The width of the remaining portion of the distal end is 4·7 cm and I estimate the original width here as at least 5·8 cm.

As will be apparent from the photographs, the shaft tapers smoothly on the medial side from the distal end, and the lateral side from the "flange".

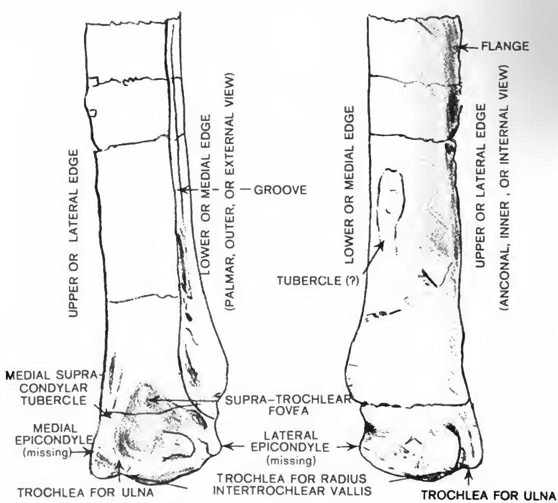


FIG. 1—Nomenclature of bird osteology as it relates to this bone.


FIG. 2—A: Outer or palmar view of distal portion of humerus. B: Inner or anconal view of humerus. C: Outer, or palmar view of humerus.

There is no indication of the expansion of the shaft which precedes the proximal end or head. This expansion is normally found in birds, and its absence indicates that at least one quarter of the bone is missing. The complete bone would have had a *minimum* length of 60 cm.

From the distal end, for a distance of about 18 cm a groove runs along the medial edge of the shaft, whence it continues as a crack. It is difficult to see exactly where the groove ends and crack begins, because of matrix in both, and the absence of small pieces of the surface. The bone wall, where it can be measured, is 1 mm in thickness; rather thinner than might have been expected in a bird of this size. The measurement was taken towards the proximal end of the remaining shaft. A fragment from the humerus of an Albatross, *Diomedea* sp. (probably *epomophora*) has a thickness of 2 mm at approximately the corresponding position on the bone.

The supra-trochlear fovea is shallow about 1 mm from the highest portion of the trochlea for the radius, but only about 3 mm at the lowest portion of

the latter. It is ovoid, rounded at the lower end, but tapering towards the upper end—"pear shaped".

The trochlea for the radius is 33 cm in length, and has an ancient abrasion on the distal end. The maximum width is 1·55 cm. The intertrochlear vallis is very shallow, only 2 mm. The trochlea for the ulna is imperfect, a portion being missing distally. It is 2·25 cm in length, and the surviving portion has a maximum width of 1·5 cm. The medial supra-condylar tubercle is shallow and the medial epicondyle is missing. The lateral epicondyle is also missing. The groove on the medial side, mentioned above, is still largely filled with matrix, but near its beginning, distally, is approximately 7 mm deep (it is difficult to measure with complete accuracy), and 3·5 mm wide. It tapers to 3 mm further along the shaft. The lateral "flange" is 20 cm in length from the distal beginning to its finish along the shaft. Its original width is unknown. In section the shaft is something like a rounded triangle toward the distal end of the humerus, which contains the "flange" and groove (Fig. 1, 2).

The inner (or anconal) side is somewhat less complicated. It has a generally flat appearance as far along the shaft as the "flange" ran. Near the medial edge, 7·7 cm from the distal end, appears the base of a long tubercle (?) 3·95 cm in length, 1·05 mm in maximum width. The whole of the distal area on this side is very flattened until one reaches the moderately pronounced trochlea for the radius. Otherwise the inner side proceeds in a smoothly flowing surface, proximally along the shaft. Some of the flattening is possibly due to compression, but this does not appear to be great, and certainly most of the shaft is rounded in a manner normally found in birds (Fig. 2B).

## DISCUSSION

Despite considerable differences, the humerus resembles that of a Pelican more closely than that of any other bird available for comparison. Because of this fact, and its enormous size, I consider it very probable that the bone belongs to a member of the Odontopterygia. Howard (1957, p. 21) placed the bony-toothed birds in the order Odontopterygiformes, but Howard and Warter (1969) follow Lambrecht (1933), Wetmore (1960) and Brodkorb (1963) in placing them in the order Pelicaniformes; Sub-order Odontopterygia, and I follow this classification.

One other specimen of the sub-order has been described from New Zealand, *Pseudondontornis stirtoni* Howard & Warter. This, an incomplete cranium and mandible, and associated femur was found in a concretionary boulder on Motunau Beach, 22 km (14 miles) north-east along the coast from the locality of the humerus under discussion. *P. stirtoni* is probably of Waitotaran (Upper Pliocene) age, but could possibly have been derived from deposits as old as Lower Miocene. Howard and Warter considered, from their study of the cranium, that *P. stirtoni* is smaller than the other *Odontopterygia*. The referred femur, badly crushed and eroded, is only 12·95 cm in lengh. A recent, but extinct, large Pelican from New Zealand, *Pelecanus conspicillatus novaezealandiae* Scarlett (1966) has a femoral length varying from 12·2 to 12·8 cm and the humerus measures from 35·5 to 35·7 cm in length.

The genera and species of the sub-order Odontopterygia Spulski, 1910 are:

Family ODONTOPTERGIDAE Lambrecht, 1933

Genus Odontopteryx Owen, 1873

**Odontopteryx toliapicus** Owen, 1873. Lower Eocene. From the London Clay of Sheppey Isle, Kent.

Family PSEUDONTONORTHIDAE Lambrecht, 1933

Genus Pseudontornis Lambrecht, 1930

**Pseudodontornis longirostris** (Spulski, 1910). The locality and age of the holotype are unknown. A fragment of jaw from the Miocene, South Carolina was referred to this species by Hopson (1964).

**Pseudontornis stirtoni** Howard & Warter, 1969. Probably Waitotaran (Upper Pliocene), but possibly as old as Lower Miocene. East Coast, South Island, New Zealand.

Genus Osteodontornis Howard, 1957

**Osteodontornis orri** Howard, 1957. Upper Miocene, California. A second specimen, also Upper Miocene, California was described by Howard and White in 1962, and another fragment from Californian Miocene was referred to this genus by Howard in 1966

For this discussion the most important bird is *Osteodontornis orri,* as the part skeleton of the first specimen included partial humeri, as well as other wing bones. As with the New Zealand humerus, only the distal ends and incomplete shafts were present. Howard describes the humeri as "bone distorted, but appearing slender except for large distension on inner border of anconal surface near distal end; this may be due entirely to distortion. Ectepicondylar prominence large". The length of the available portion of the humerus is 59·3 cm, and the distal breadth (width in my terminology) approximately 3·4 cm. From the photographs published by Howard (1957) the better-preserved humerus of *O. orri* bears some resemblance to that of the New Zealand bone, although the shaft is straighter than that of the latter, and no "flange" is visible. Howard's *minimum* estimate for a single wing of *O. orri,* based on the measurable bones and feather impressions, is 197 cm (6 ft 5½ in.) and as she points out, this could have been considerably larger.

Unfortunately, the figures of *Osteodontornis orri* are not clear enough to permit of a more detailed comparison between the two humeri.

The length of a humerus of a Royal Albatross, Canterbury Museum AV15,669, is 42·5 cm and the combined length of the (dissociated) bones, humerus, ulna, carpo-metacarpus and index digit, gave a length of 113·5 cm.

In this skeleton, the ulna is approximately the same length as the humerus, and the combined length of carpo-metacarpus and index digit roughly two-thirds the length of the humerus. The holotype of *Pelecanus conspicillatus novaezealandiae* has a humerus length of 35·5 cm, and the ulna is longer, 38·0 cm. The carpometacarpus is 16·7 cm in length. The index digit is unknown, but assuming that combined with the carpo-metacarpus it was about two-thirds of the humerus, the Pelican would have had a total wing length of 99 cm without feathers.

Using the same formula, and taking the minimum estimated length of the complete humerus as 60 cm, we can arrive at a minimum wing length, minus feathers, of 160 cm for the New Zealand bird. The longest wing feather,

which Dr Howard added to the bone lengths to get her estimated minimum length of 197 cm was 30 cm, so that the wing of the New Zealand bird would have been very little shorter than that of *Osteodontornis orri*.

Of course, one must allow a margin of error in these estimates, as we are working with incomplete material, but it is clear that both birds had an enormous wingspread.

It is unwise to give a name to the bird represented by this partial humerus, until more bones, preferably including a cranium and beak, are found, but a description seemed necessary, as the available material of the "bony-toothed" birds is still so scanty that every scrap of evidence must be presented for the information of other workers.

ACKNOWLEDGMENT

The photographs were taken by Mr David Jones, Department of Geology, University of Canterbury; the white appearance in the photographs is due to coating of the bone by ammonium chloride before photography.

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# NOTES ON THE TAPHONOMY AND PALEOECOLOGY OF NEW ZEALAND TERTIARY SPATANGOIDA

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(*Received 9 March 1971*)

ABSTRACT

The distribution of spatangoid echinoids and molluscs in New Zealand Tertiary formations is anomalous; formations relatively rich in spatangoids are poor in molluscs and vice versa. The absence of Mollusca from spatangoid-rich formations is attributed to diagenetic dissolution of aragonitic shells while the stabilisation of spatangoid magnesium calcite is attributed to incongruent dissolution.

Irregular bevelled perforations in some spatangoid tests are attributed to predation by tonnacean gastropods such as *Echinophoria* and large *Austrosassia*. New Zealand Tertiary spatangoids are of little use as depth indicators as those genera still living have large depth ranges.

## INTRODUCTION

Following a review of New Zealand fossil spatangoids by one of us (Henderson, in press) the anomalous distribution of Tertiary spatangoids with respect to molluscs became apparent from discussions between Beu and Henderson, and the third author was invited to comment from his specialist knowledge of carbonate diagenesis. The study of taphonomy, which is concerned with changes in faunal assemblages between their inferred complements in life and their stabilisation in the fossil record, is in its infancy. The loss of aragonitic shells from certain formations as documented here may well prove to be a general and important taphonomic process; it has recently been reported from an Oligocene oyster community from North Carolina, U.S.A. (Lawrence 1968, p. 1324) and from rocks of the Germanic Muschelkalk basin (Seilacher 1971, p. 16 *et seq.*).

The paleoecological notes follow from inferences on the nature of predators which drilled perforations in a small percentage of spatangoid tests and from an assessment of the usefulness of spatangoid taxa in depth paleoecology; they are the responsibility of Beu and Henderson

Age data are given in terms of New Zealand local stages, with the recent modifications of Scott (1972).

## RELATIVE ABUNDANCE OF MOLLUSCS AND SPATANGOIDS IN NEW ZEALAND TERTIARY FAUNAS

In New Zealand Tertiary rocks spatangoids are most commonly preserved in shallow water sandstones and detrital limestones. Specimens have been collected from over 50 individual formations but a few are especially important for spatangoids (Table 1).

The total number of catalogued New Zealand spatangoid specimens in the collections of museums, universities and the New Zealand Geological Survey, with the exception of a few very poorly preserved specimens, is approximately 830. Specimens from the 10 formations listed in Table 1 therefore account for over 40% of the total collection of New Zealand Tertiary spatangoids.

Tertiary molluscan faunas, in comparison, are much richer and the total collection housed in reputable institutions is well in excess of 100,000. The ten formations listed in Table 1 are significantly poor in their molluscan complement and together account for less than 1% of the total New Zealand Tertiary molluscan collection. Only one formation, the Pakaurangi Formation, has yielded abundant molluscs and a significant number of spatangoid specimens (16). In most of the facies characteristically rich in Mollusca (such as shelf blue-grey siltstones) spatangoids are rare or have not been collected.

It can be concluded that in general a strong disparity exists between formations relatively rich in molluscs and those relatively rich in spatangoids. This relationship may be in part due to ecological reasons but the following points argue against it being entirely so:

1. In all 10 formations listed in Table 1, the only well preserved Mollusca are calcitic bivalves (mainly Pectinidae, Limidae, Ostreidae and Anomiidae)

TABLE 1—Numbers of Specimens from Important Spatangoid-bearing Formations in which Molluscs are Relatively Rare. Numbers are approximate in some instances because many specimens are inexactly located; the total can be considered as a minimum value

| Formation and Locality | Age | Number of Specimens |
|---|---|---|
| Island Sandstone, northern Westland | Ab | 103 |
| Curiosity Shop Greensand, central Canterbury | Lw | 48 |
| Abel Head Formation, north-western Nelson | Ld–Lw | 47 |
| Caversham Sandstone, eastern Otago | Po–Pl | 40 |
| Cobden Limestone, northern Westland | Lwh | 35 |
| Otekaike Limestone (Maerewhenua Member), South Canterbury | Lw | 19 |
| Glen Massey Formation (Te Kuiti Group), West Auckland | Lwh | 14 |
| Aotea Sandstone (Te Kuiti Group), West Auckland | Lwh–Ld | 14 |
| Mount Brown Limestone, North Canterbury | Pl | 14 |
| Waihao Limestone, South Canterbury | Ld–Lw | 13 |
| | Total | 347 |

and gastropods (Epitoniidae), and as reported earlier by one of us (Beu 1967, p. 241), the aragonitic layers of anomiid shells are absent. Mollusca which construct their shells of aragonite are uncommon or rare and where present are usually preserved as moulds. Brachiopods (also calcitic) occur with spatangoids in some of these formations (e.g., *see* Gage 1957, p. 51). They are also the main rock units from which fossil bones (calcium phosphate) of Cenozoic marine animals (e.g., penguins and whales) have been collected in New Zealand. The absence of aragonitic shells in their fossil complement is most unlikely to reflect a biological cause.

2. All the well preserved Mollusca are epifaunal with infaunal elements almost totally lacking.

3. There is no reason to suspect from present day ecology that molluscs and spatangoids are mutually exclusive. Faunal association data are lacking for many spatangoid species but reviews of marine bottom communities by Jones (1950) and Thorson (1957) suggest that spatangoids and molluscs are common community associates.

4. Evidence from predator borings (*see below*) shows that carnivorous gastropods have preyed on a small percentage of spatangoids prior to fossilisation. As most tests are still complete we presume that boring was shortly before fossilisation and would expect to find the predators fossilised with their prey on at least some occasions.

We conclude that aragonitic molluscan shells have been selectively removed from most formations in which spatangoids form a high proportion of the preserved fauna. As discussed below, selective removal finds theoretical support from considerations of calcium carbonate diagenesis. The converse relationship of a low spatangoid density in formations rich in molluscs is not readily explained but the following points are relevant:

1. Collection failure is doubtless a contributory factor and spatangoids have been discriminated against where fossils are abundant because of their limited value for age determination.

2. Spatangoid tests are much more fragile than those of the Mollusca. In at least some cases. molluscs in shell beds have been concentrated by erosion with the removal of the formerly enclosing sediment; spatangoid tests are unlikely to have survived this process. Pronounced compaction typical of fine grained sediments, such as the silt and mud in which molluscs are commonly abundant but spatangoids rare, may well have resulted in crushing and partial disarticulation of imbedded spatangoid tests. These when exposed at outcrop surfaces would rapidly disintegrate and their likelihood of collection decline accordingly. This is probably why spines of the spatangoid *Echinocardium cordatum* (Pennant) are common in Tainui Shellbed, Castlecliff (Castlecliffian), a rich molluscan shellbed where even fragmentary tests of *Echinocardium* are extremely rare. Finally, the markedly greater fragility of spatangoid tests means they are much more difficult to remove from compacted mudstone than are hard molluscan shells, especially if the spatangoids have been crushed during compaction.

The possibility that New Zealand Tertiary spatangoids had a distinct preference for coarser grained sediments when alive, such as the sandstones and detrital limestones from which they are presently known in greatest abundance, cannot be discounted. Certain extant species such as *Meoma*

*ventricosa* (Lamarck), *Plagiobrissus grandis* (Gmelin) and *Paraster floridiniensis* Kier and Grant are known to favour sandy substrates (Kier and Grant 1965).

*Mineralogical considerations*

Invertebrate carbonate skeletons are composed of the minerals calcite (arbitrarily defined as containing less than 4 mole percent $MgCO_3$), aragonite, or any of a range of magnesium calcites (arbitrarily defined as containing greater than 4 mole percent $MgCO_3$). While biogenic aragonite and magnesium calcite form the dominant skeletal minerals in modern warm shallow marine waters, they are metastable carbonate phases, and in time are eliminated or replaced by stable calcite or, more rarely, some other mineral. The solubility of aragonite and magnesium calcite in natural waters is considerably higher than that of calcite, and for magnesium calcite the solubility increases with increasing $Mg^{2+}$ content (Chave *et al.* 1962). Under natural near-surface conditions a general stability sequence among the three carbonate minerals is: calcite > aragonite > magnesium calcite (Stehli and Hower 1961). The primary mineralogy of carbonate tests may therefore be an important factor in affecting fossil preservation and controlling subsequent diagenetic events.

Chave *et al.* (1962) suggested that selective dissolution, and therefore total loss, of metastable carbonate skeletons may occur on the sea floor. Several authors, including Stehli and Hower (1961), Friedman (1964), Dodd (1966), and Land (1967), cite evidence for the stabilisation of metastable carbonate minerals within sediments or sedimentary rocks after a short period of burial or uplift. In the case of magnesium calcites the stabilisation reaction normally involves a process of incongruent dissolution (Land 1967), whereby $Mg^{2+}$ is lost to solution, yielding a replacement product of calcite. The process is texturally non-destructive, so that original fossil outline is faithfully preserved and test microstructures remain completely intact. The stabilisation of aragonitic skeletons, however, commonly involves complete dissolution of the mineral which, depending on the degree of lithification of the enclosing sediments, may or may not leave a mould. In this case evidence of former aragonitic shells may be either entirely lost or drastically reduced. Thus although magnesium calcitic phases are more metastable than aragonitic ones, they are more likely to be preserved following diagenetic stabilisation reactions.

It is considered that the anomalous distribution of molluscan and spatangoid faunas in New Zealand Tertiary rocks may be explained by these diagenetic relationships. All echinoids construct skeletons of metastable magnesium calcite (Dodd 1967). Specific information for spatangoids given by Weber (1969) shows an average value of 11·3 and 6·28 wt percent $MgCO_3$ for tests and spines respectively. Analysis of the magnesium content of the skeletons of echinoids, calcareous red algae and bryozoans from the Aotea Sandstone and other Lower Tertiary formations in south-west Auckland by one of us (Nelson in prep: Sedimentology and stratigraphy of the Te Kuiti Group in the Waitomo County, South Auckland Land District. Ph.D. thesis, University of Auckland) shows that the original magnesium calcite of these organisms has been replaced by stable calcite. Molluscs have skeletons of metastable aragonite or a variable combination of aragonite and stable calcite;

a few bivalves, and most notably the oysters, are virtually entirely calcitic (Dodd 1967). The general absence of originally aragonitic molluscs and other aragonitic skeletons, with the restriction of the fossil fauna to calcitic or originally magnesium calcitic shells, is consistent with the view that metastable aragonitic phases have been lost from the spatangoid-rich lithologies.

Judging from the literature, it is clear that the time of stabilisation of the metastable carbonate minerals is unpredictable and, depending on circumstances (*see* for example, conditions given in Dodd 1966; Land 1967; and Weber 1969), may occur within the water mass, at the sediment–water interface, or at any subsequent stage in the diagenetic history of a sediment or rock. Considering the range of lithologies represented, the time of stabilisation undoubtedly varied and must remain conjectural, although a few data exist for Lower Tertiary rocks (Te Kuiti Group) in south-west Auckland. Here a variety of evidence suggests the calcareous lithologies were cemented by calcite derived from intrastratal solution of tests soon after deposition (Ballance and Nelson 1969; Nelson in prep. *see above*). This suggests that in these cases stabilisation of aragonite and magnesium calcite skeletons was also a very early diagenetic reaction, for otherwise aragonitic fossils, casts or moulds would be common, and the cemented rocks might be expected to contain a high magnesium content and perhaps be dolomitised. The average magnesium content of the acid-soluble fraction of 48 analysed sandstones and limestones in the Te Kuiti Group is only 0·35% and dolomite has not been positively identified (Nelson in prep. *see above*). The vast majority of $Mg^{2+}$ released during incongruent dissolution was apparently either lost from the system or, more probably, absorbed by clay minerals before lithification and while permeability was high. In particular the especially reactive nature of echinoderm magnesium calcite is consistent with its very early stabilisation (Land 1967).

Different aragonitic skeletons show varying susceptibility to replacement by calcite, and stabilisation may therefore occur over a long period of time. Evidence from the Aotea Sandstone (Nelson in prep. *see above*) suggests that dissolution of aragonite may have occurred before, during and following cementation of the sediment. As suggested above, the solutional loss of considerable amounts of aragonitic material is believed to have been most important during the first of these stages. Evidence for such a pre-cementational loss is perhaps provided by the complete absence of either a mould or cast of the original aragonitic layer of certain bimineralic bivalve shells (e.g., the anomiid *Pododesmus*). Rare aragonitic bivalves survived early stabilisation reactions, as shown by the occurrence of occasional moulds of aragonitic bivalves (e.g., of *Cucullaea* and *Panopea* in the Aotea Sandstone). The fact that some moulds are distorted suggests, in this case, that aragonite dissolution operated before complete cementation and while sediments were still plastic. On the other hand, undistorted moulds that are entirely empty of either cement or sediment suggest post-cementational leaching of aragonitic shell material by meteoric water following uplift and subaerial exposure of the rocks. A close inspection of these moulds showed that some contain small quantities of a white chalky material which, upon X-ray investigation, proved to be aragonite.

The persistence of metastable carbonate minerals in New Zealand rocks is at present being investigated by one of us (C.S.N.). Preliminary results suggest that some of the oldest aragonite is preserved in those marine sediments in which locally reducing conditions were perhaps important as evidenced by the proximity of coal measures in the stratigraphic column and the moderate amounts of organic matter and the abundance of glauconite and pyrite in the sediments. Under these conditions the oxidation of the organic matrix of shells is inhibited, thereby isolating the crystals of aragonite from water and protecting them from dissolution (Kennedy and Hall 1967). This possibly explains why certain formations (e.g., Wharekuri Greensand, North Otago) show the expected association of spatangoids and a rich molluscan fauna.

### *Conclusion*

We conclude that the paucity of molluscs relative to spatangoids in many New Zealand Tertiary rocks is due to the early diagenetic dissolution of considerable amounts of molluscan aragonite from the rocks in which spatangoids are relatively common; stabilisation of the more metastable magnesium calcite in spatangoid tests also occurred soon after deposition, but involved the texturally nondestructive process of incongruent dissolution which permitted preservation of the echinoid fauna.

## PREDATION OF SPATANGOIDS

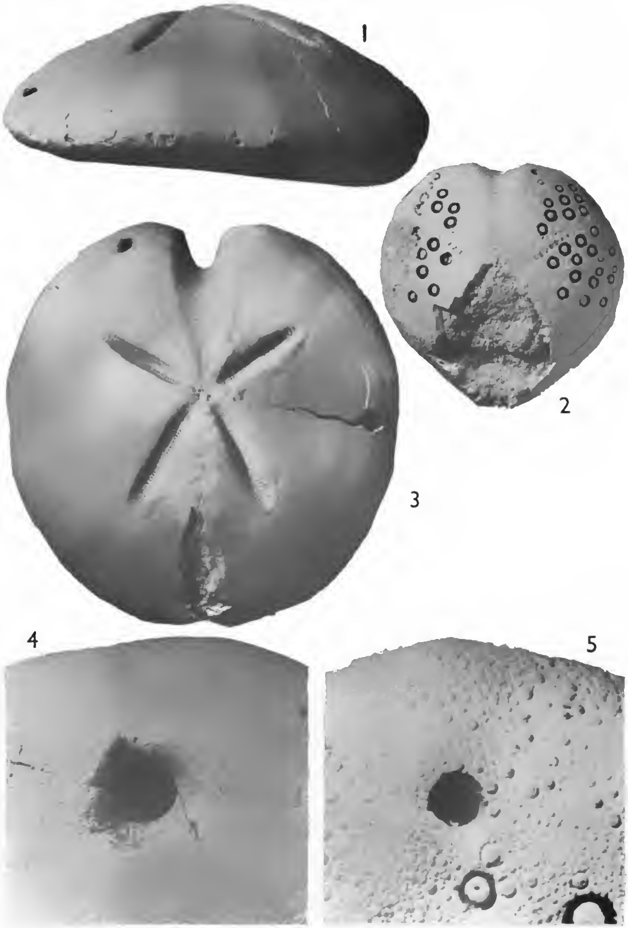
Several large spatangoids in collections of New Zealand echinoids examined by one of us (R.A.H.) have perforations in the test. The perforations have irregularly bevelled margins with the bevel sloping outwards to the outer test surface, so that the outer aperture of the perforations is roughly polygonal; at the inner test surface, the narrow ends of the foramena are nearly circular. Table 2 shows the position of such perforations in New Zealand spatangoids examined, and their dimensions. Some tests listed in Table 2 are incompletely perforated, and bear an irregularly bevelled pit. Two completed perforations are shown in Fig. 1–5.

Table 2 shows that on all but *Cyclaster,* the perforation is most commonly adjacent to the ambitus in the interambulacral plate series, on the aboral surface. However, in *Cyclaster posita* (Hutton) from "Curiosity Shop", Rakaia River (Waitakian), the three perforated specimens all bear perforations on the oral surface, in the plastron or the first ambulacrum. Apart from position, the perforations are similar. The different positions seem to indicate a slightly different cause for perforations in *Cyclaster* from those in other spatangoids.

Naticacean, tonnacean and muricacean gastropods are well-known carnivores, preying on invertebrates with hard skeletons, such as echinoderms and molluscs (Carriker and Yochelson 1968, p. 33). Of these, naticacean and muricacean borings are abundant, well described, and easily differentiated; Naticacea drill neatly circular holes with a gently sloping, slightly concave bevel, leaving a central boss in incomplete holes, whereas Muricacea drill holes which are similar but more steeply bevelled and lack the central boss

TABLE 2—Positions of Predator Borings in New Zealand Tertiary Spatangoid Tests. Catalogue numbers are of collections from Department of Geology, University of Otago (numbers prefixed by OU), New Zealand Geological Survey (EC), Canterbury Museum (zfe), and Department of Geology, University of Auckland (E). S36/f413 and similar numbers are localities in the National Fossil Record, showing the number of the one-mile topographic map sheet (NZMS 1), and the number of the locality within the sheet. Diameters of borings are maxima, shown in mm. A dash in the internal dimension column shows lack of penetration of test. Stage symbols: Ld = Duntroonian; Lw = Waitakian; Po = Otaian; Pl = (revised) Altonian

| Catalogue number | Echinoid name | Position of Boring | Diameter of boring Internal | External | Locality | Age |
|---|---|---|---|---|---|---|
| E236 | Pericosmus crawfordi (Hutton) | Aboral surface, interamb 3, adjacent to ambitus | 3.7 | 6.0 | S36/413, Ngapara | Lw-Po |
| zfe 38 | as above | Aboral surface, close to boundary between interamb 1 and amb 3 | - | 3.1 | S82/511, Curiosity Shop, Rakaia R. | Lw |
| E315 | Pericosmus sp. nov. | Aboral surface, interamb 2, adjacent to ambitus | - | 2.9 | N28/874 Pakaurangi Point Kaipara Harbour | Pl |
| OU.4725 | Lovenia tuberculata (Zittel) | as above | 1.5 | 2.0 | "Puponga, Nelson" | Ld-Pl? |
| zfe 199 | Eupatagus formosus (Zittel) | Aboral surface, interamb 3, adjacent to ambitus | - | 2.2 | No information | |
| EC.671 | Taimanawa pulchella (Henderson and Fell) | as above | 6.0 | 7.1 | No information | |
| OU.8548a | Lambertona lyoni (Hutton) | Aboral surface, interamb 1, halfway from apical system to ambitus | 2.1 | 3.5 | Hanging Rock, Opihi River | Ld-Lw |
| zfe 41 | Cyclaster posita (Hutton) | Oral surface, on plastron near its junction with labrum | - | 5.2 | S82/511, Curiosity Shop, Rakaia R. | Lw |
| zfe 44 | as above | as above | 3.1 | 4.9 | as above | |
| zfe 47 | as above | Oral surface, ambulacrum 1, adjacent to the labrum | 3.9 | 5.0 | as above | |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(Carter 1968, p. 38, fig. 4). Holes bored by *Octopus* in shells figured by Carter (1968, pl. 1, fig. 8, 9) are quite similar to the perforations in New Zealand Tertiary spatangoid tests, but are much smaller, still less regular, and have no bevel. The perforations in spatangoid shells are probably gastropod boreholes, but are not likely to have been drilled by Naticacea or Muricacea.

Tonnacea are becoming increasingly well known as predators on echinoderms. Among the Cymatiidae, *Charonia tritonis tritonis* (Linnaeus) is an active predator on the coral-eating starfish *Acanthaster planci* (Linnaeus) (Chesher 1969a; and many other recent references), and in New Zealand *Charonia lampas capax* Finlay and *C. lampas rubicunda* (Perry) "feed spasmodically on a variety of echinoderm species" (Laxton 1970, p. 133). Cassidae are even better known predators, apparently mainly on echinoids. Many writers have recorded the feeding of Caribbean Cassidae (*Cassis madagascariensis*; *C. tuberosa* (Linnaeus); *Phalium* (*Tylocassis*) *granulatum* (Born)) on a variety of echinoids (*Diadema antillarum* Philippi, *Lytechinus variegatus* (Lamarck), *Clypeaster rosaceus* (Linnaeus), *Plagiobrissus grandis* (Gmelin), *Tripneustes esculentus* (Leske), *Mellita quinquiesperforata* (Leske), *Meoma ventricosa* (Lamarck)), including the spatangoids *Meoma* and *Plagiobrissus* (Lyman 1937; Foster 1947; Moore 1956; Schroeder 1962; Abbott 1968, pp. 16, 17; Chesher 1969b, p. 94). Descriptions of borings are as follows: ". . . an area of about 25 mm in diameter had been cleared of spines. In this area there was a neatly drilled hole through the thin test. One specimen . . . has a hole 9 mm in diameter in the antero-lateral edge" (Moore 1956); "The holes through sea urchin tests made by *Cassis* at Bimini are more or less circular and from 4 to 6 mm in diameter. There seems to be no special area on the test which is pierced. An acid appears to be used at least in part, to make the hole, for the edge is smoothly rounded and concave all round except at each joint between plates of the test where there is either an irregular groove or a lamina" (Abbott 1968, p. 17, quoting notes by Robert Robertson). Schroeder (1962) gave an excellent set of figures of *Cassis tuberosa* catching *Diadema antillarum*. Abbott (1968, p. 16, pl. 2) figures a hole drilled in a test of *Lytechinus variegatus* (Lamarck) by a specimen of *Cassis tuberosa* (Linnaeus) when both were kept in an aquarium in the Bahamas. The edges of the hole are roughly pentagonal in outline, the angles being at juctions of plates of the test, and are decidedly ragged. There is a slight bevel around the margins but the inner edge is as irregular as the outer. It is 5·5 mm in diameter. The boring is thus similar to those in New Zealand spatangoids, but lacks the broad bevel and circular inner edge of New Zealand examples. The description by Abbott (1968, p.

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FIG. 1–5 *Opposite*

1, 3, 4—*Pericosmus crawfordi* (Hutton), showing boring attributed to a tonnacean gastropod. S36/f413, Ngapara, near Oamaru (Waitakian–Otaian); E236, Auckland University Geology Department. Fig. 1 lateral, and Fig. 3 aboral views, × ⅔; Fig. 4 boring enlarged, × 3.

2, 5—*Lovenia tuberculata* (Zittel), showing boring (at top right) attributed to a tonnacean gastropod. "Puponga, Nelson" (? Abel Head Formation, near Cape Farewell; Duntroonian to Altonian); OU 4725, Otago University Geology Department. Fig. 2 aboral view, × 1; Fig. 5 boring enlarged, × 3.

17) quoted above, however, indicates that broader bevelled edges and more smoothly rounded outlines occur in other Caribbean examples. The roughly pentagonal outline, with angle at the junction of the plates of the test, and raggedness of outline due to pulling away of fragments of the coarsely crystalline echinoid test, are particularly similar to those of New Zealand fossil examples, and we conclude that the borings in New Zealand Tertiary spatangoids were made by Tonnacean gastropods. The large number of records of boring in echinoids by Cassidae, summarised above, and total lack of records of boring by Naticacea or Muricacea (both of which normally feed on bivalves) help confirm that the borings in New Zealand spatangoids were not made by Naticacea or Muricacea.

Tonnacea that could have been preying on New Zealand spatangoids listed in Table 2 (age range: Duntroonian–Altonian) are Cymatiidae, such as the large species of *Austrossassia* of the *zealta* lineage (Beu in prep.: New Zealand Cenozoic and Recent Mollusca of the family Cymatiidae, with a review of world genera) and Cassidae, such as the several species of *Echinophoria*. It is not possible to assign them more closely to individual taxa. The two positions of boring in spatangoids—on the oral surface of *Cyclaster* and the aboral surface of all other genera—suggest that two different species of Tonnacea may have been responsible for the borings. It is possible, for instance, that one group of borings was made by *Echinophoria* and one by *Austrosassia*. It is also possible, however, that the relatively high, inflated form of *Cyclaster* made it easier for the same predator to grip *Cyclaster* in a different way from the way it gripped other spatangoids. It is unlikely that these points can be resolved, because predatory gastropods are not often preserved with the spatangoids.

While the site of boring is usually random for Cassidae (Abbott 1968, p. 17), it tends to be near the anterior end of the test in New Zealand spatangoids. We suspect that this tendency is common, because an echinoid held at the anterior end is unable to turn, and thus cannot escape its predator; also, spatangoid spines are directed backward from the anterior, so that the gastropod would have less trouble pushing them aside when holding a spatangoid at the anterior end. However, variable position of boring may reflect merely taxonomic diversity of borers.

## DEPTH PALEOECOLOGY

Present day spatangoid taxa are commonly tolerant of a wide range of depths compared with other benthonic marine invertebrates. Many species range from inner neritic to bathyal depths and some range from neritic to abyssal depths (e.g., *Brissopsis mediterranea* Mortensen, 46–3200 metres). Table 3 gives the known depth ranges for living genera and the one living species which are represented in the Cenozoic fossil record of New Zealand. Data in Table 3 are largely from Mortensen (1951).

It is concluded that New Zealand Cenozoic spatangoids are of little use for estimating the depositional depth of beds in which they are found.

TABLE 3—Depth Ranges of Living Genera and One Living Species of Spatangoids that Occur as Cenozoic Fossils in New Zealand

| Genus | Depth of Occurrence (m) |
|---|---|
| *Schizaster* (s.s.) | 5–365 |
| *Paraster* | *c.*20–900 |
| *Diploporaster* | shallow water |
| *Prenaster* (= *Protenaster*) | 12* |
| *Lovenia* | subtidal–930 |
| *Echinocardium cordatum* (Pennant) | littoral–230 |
| *Brissus* (s.l.) | littoral–240 |
| *Brissopsis* | 7–3200 |
| *Meoma* | littoral–200 |
| *Eupatagus* (s.s.) | 10–600 |
| *Cyclaster* | 210–475 |
| *Taimanawa* | 260* |
| *Maretia* | subtidal–150 |
| *Pericosmus* | 18–486 |

*One record only.

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# *GLOBIGERINOIDES* FROM ESCORNEBÉOU (FRANCE) AND THE BASAL MIOCENE *GLOBIGERINOIDES* DATUM

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## Abstract

The first evolutionary appearance of *Globigerinoides trilobus primordius* (the *Globigerinoides* datum) has been taken to correspond closely with the base of the Aquitanian Stage at its stratotype, Saucats, south-western France. But elsewhere in Aquitaine basin (Escornebéou) the taxon occurs in strata with miogypsinids that are antecedent to those in the Aquitanian stratotype. Biometric comparison of *G. trilobus primordius* from these localities supports the miogypsinid data. The Escornebéou population is probably more primitive than those from Saucats: heights of apertures and of the spire are greater and chamber overlap is smaller. The test is less compact. As a precise criterion for the base of the Aquitanian Stage, as judged from the stratotype, the appearance of *G. trilobus primordius* is unreliable. Its first evolutionary appearance precedes this horizon.

## Introduction

*Globigerinoides trilobus primordius* was reported by Butt (1966) from Escornebéou, on the southern flank of the Aquitaine basin, in strata that he considered to be Chattian. This occurrence bears on the nature, position and utility of the *Globigerinoides* datum (Bandy 1964; Scripps 1969; Blow 1969) and on the early history of Neogene *Globigerinoides*. The Escornebéou record is significant because the state of nepionic development in associated miogypsinids (Butt 1966) provides strong independent evidence that the strata are inferior to those on the north-eastern flank of the Aquitaine basin, near Saucats, that constitute the stratotype of the stage. This relationship, if correct, raises doubt about the close coincidence of the first evolutionary appearance of *Globigerinoides* (the "*Globigerinoides* datum") with the base of the Aquitanian stratotype. Blow (1969) reported that the divergence of *G. trilobus primordius* from an ancestral *Globigerina* occurred at an horizon "extremely close to the horizon represented by the base of the lectostratotype of the Aquitanian in the valley of Saucats".

Several interrelated questions that arise from these data may be resolvable from comparison of the Escornebéou and Saucats populations. Do these populations in fact belong to the same lineage? If so, does their morphology give clues to their relative evolutionary positions and possible stratigraphic order?

This paper compares *Globigerinoides* in a re-collection (F100718 in N.Z. Geological Survey micropaleontology laboratory collections) of locality 9 of Butt (1966, fig. 3) from Escornebéou with those in re-collections of localities

15a (= F100700) and 19b (= F100706) of Drooger *et al.* (1955, fig. 2, 4). The latter collections are from strata in the type region of the Aquitanian Stage in the vicinity of Saucats. Locality 15a is from the section at Moulin de Bernachon which constitutes part of the Aquitanian stratotype; the horizon is classified as middle Aquitanian by Drooger *et al.* (1955, distribution chart). They note that the fauna includes *Miogypsina tani* Drooger. Locality 19b is 3 km east of the stratotype. *M. gunteri* Cole is present. Drooger *et al.* classify the locality as lower Aquitanian.

Specimens (24 from each collection) were obtained from residue retained on a sieve with nominal apertures of 0·152 mm diameter. Only those in which the final chamber ($n^{th}$) was intact and the rim of the primary aperture linked the $(n-1)^{th}$ and $(n-3)^{th}$ chambers were studied. Variates are continuous linear measurements (Fig. 7–10) and reflect chamber expansion (×1–×4), sizes of the primary and last-formed supplementary aperture (×5–×8), geometry of the test (×9–×12) and shape of the final chamber (×13–×14). Note that, because supplementary apertures develop late in ontogeny, only a restricted age segment of the population is sampled.

## ONE LINEAGE OR TWO?

The phylogenetic problem arising from the work of Butt (1966) is the occurrence of *Globigerinoides trilobus primordius* in strata at Escornebéou that may be inferior to those at Saucats in which its first evolutionary appearance was reported by Blow (1969). One solution to this conundrum is that there has been confusion of populations from separate lineages. Little is known about the earliest phylogeny of Neogene *Globigerinoides*. Moreover, in the history of Cenozoic planktonic foraminifera, acquisition of supplementary apertures is not an event unique to Neogene *Globigerinoides*. It cannot be assumed that the group is monophyletic. Patterns of variation within samples provide data on which to base an assessment of propinquity.

### *Analysis*

1. Size variation is portrayed by the largest principal component (Fig. 11) extracted from each sample covariance matrix. Of the total variance of the 14 variates in F100718 (Escornebéou) 82·4% is absorbed by this component but smaller percentages (62·4 for F100700, 55·7 for F100706) are absorbed by it in Saucats collections. The length of the first principal axis of the sample ellipsoid is greatest in F100718.

2. The second and smaller principal components in each analysis reflect shape variation (Fig. 11). This term is used in the sense of concomitant increase in some variates and decrease in others, and is indicated by change in sign of elements in eigenvectors. The second and third principal components in each

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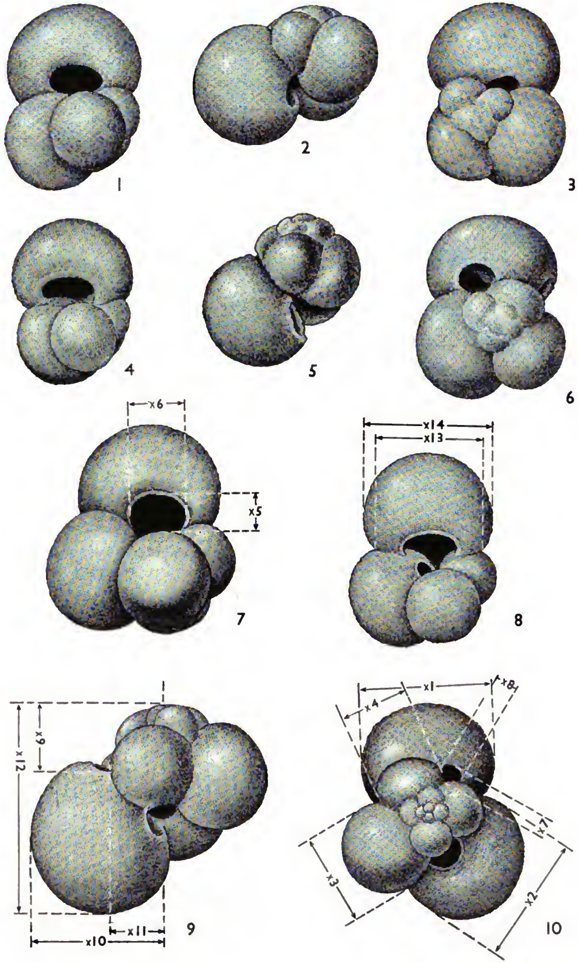
FIG. 1–10—*Globigerinoides trilobus primordius* Blow and Banner s.l. ×90.
1–3 F100700 Moulin de Bernachon, middle Aquitanian FP 2154.
4–6 F100706 Labrède, lower Aquitanian FP 2155.
7–10 F100718 Escornebéou, FP 2156.




— *R. C. Brazier, del.*

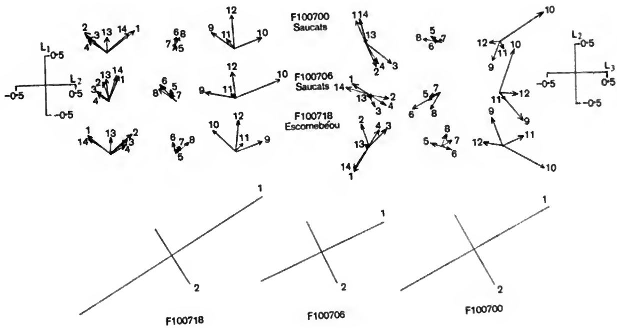



FIG. 11—(*upper*) Projections of original variate axes on to planes defined by axes for first and second (*left*) and second and third (*right*) principal components of sample covariance matrices. Varying dispositions of variate axes indicate that principal axes of the ellipsoids of scatter change in orientation from sample to sample.
(*lower*) Relative lengths of first and second principal axes. Orientation of the first axis in each sample in relative to $\times 12$, the variate with largest variance in each sample .

analysis reflect shape variation between the final and preceding chambers. This is due to irregularity in growth in final stages of ontogeny and occurs commonly in planktonic foraminiferal populations. Variation in coiling arrangement is shown by the second component for each analysis. In Aquitanian collections increase in horizontal translation ($\times 10$) is accompanied by decrease in measurements of axial height ($\times 9$, $\times 12$) and in translation of the penultimate whorl ($\times 11$). The converse occurs in F100718 (Escornebéou). Shape variation in apertures ($\times 5$–$\times 8$) is very minor in the second and third components.

3. The null hypothesis that covariance matrices are homogeneous (Seal 1964) is rejected ($\chi^2 = 838{\cdot}64$, d.f. $= 210$, $P < 0{\cdot}01$). This indicates that orientation and/or inflation (Fig. 11) of sample ellipsoids differ. Collinearity of first principal axes between pairs of ellipsoids is rejected ($P < 0{\cdot}01$) by an approximate test (Reyment 1969). Lack of collinearity in this axis indicates that, in terms of the 14 variates, the direction of growth varies from sample to sample.

4. Intensity of relationship between pairs of variates is estimated by the product-moment correlation coefficient. In F100718 all coefficients are positive and differ from zero at the 5% level. Correlation between variates is less strongly established in the Aquitanian collections. At the 5% level, horizontal translation ($\times 10$) is correlated only with $\times 1$ and $\times 14$ in F100700

and with ×1, ×2, ×13 and ×14 in F100706. Coefficients between aperture heights (×5, ×7) and variates estimating coiling geometry (×9–×12) in F100706 are mostly not significant at the 5% level.

5. Morphological affinity may also be assessed from relations among sample scatters on common axes. Bivariate plots on original variate axes show extensive overlap among samples. Typically that for F100718 encloses those for the Aquitanian collections, F100706, F100700 (Fig. 12). A principal component plot (Fig. 12) of pooled collections using the axes for the two largest components gives a similar configuration. This is a very crude portrayal in view of inhomogeneity among sample covariance matrices.

*Interpretation*

Intra-sample variability is greater in the Escornebéou collection than in those from Saucats. Principal component analyses indicate that it is primarily due to a greater scatter of specimens in the direction of growth. As such it is unlikely to be useful as a discriminator among phylogenies. It is probably a real attribute of the Escornebéou population rather than an artefact produced by depositional and post-depositional processes or sampling. A principal component analysis of a subsample of individuals from F100718 in which the number of chambers was constant (12) again showed that the largest component (reflecting size variation) accounted for over 80% of total variance. This result suggests that variation in rate of chamber expansion may be the cause.

Direction of growth is potentially valuable for discrimination among lineages. The data indicate that the first principal axes are not collinear, but do not suggest that the difference between the Saucats collections is less than those between Saucats and Escornebéou.

The presence of closely similar phenotypes, in terms of the variates considered, is indicated by the degree of overlap among scatters when plotted on common axes. This may be due to close genetic relationship but could also arise from convergence of unrelated stocks or parallelism of recently divergent lineages.

The simplest interpretation of the data is that the collections represent transects of a single lineage within which patterns of intra-sample variation are variable. Similarity of morphology of specimens among the samples is accepted as evidence of common descent.

## MAPPING THE LINEAGE

Computation of distances between sample mean vectors permits the construction of a topology from which relationship of populations in a lineage may be inferred. The configuration shown in Fig. 12 is based on computation of Mahalanobis' $D^2$ using the pooled 3-sample covariance matrix. The problem of estimating distance between populations with inhomogeneous covariances matrices was investigated by Chaddha and Marcus (1968). They found $D^2$ to be a consistent estimator of the interpopulation distance when sample sizes are equal, as in the present data.

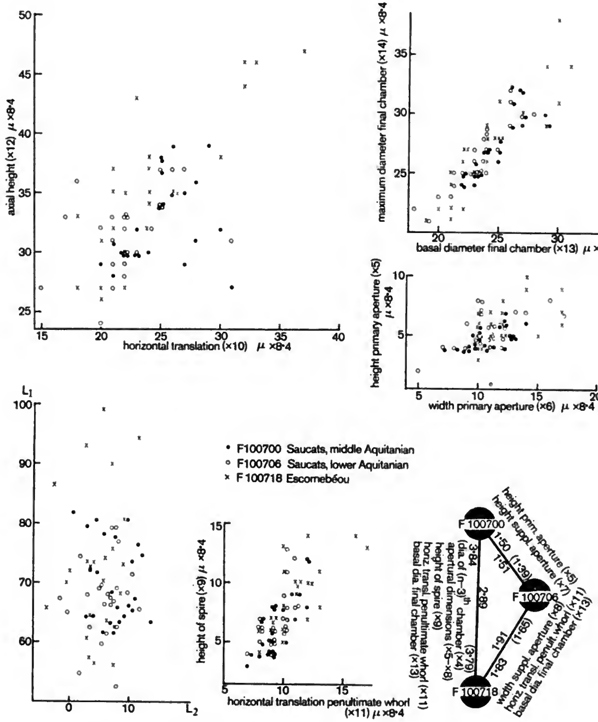



FIG. 12—(*upper three and lower centre*) Bivariate scatters for selected variates. Those for Saucats collections are more restricted than for Escornebéou, especially in the direction of growth.

(*lower left*) Projection of specimens using axes ($L_1$, $L_2$) defined by first and second principal components of covariance matrix for pooled collections. These axes absorb 81·6% of total variance. The representation of specimen topology in the 14-variate hyperspace is very crude as inequality of covariance matrices for individual samples is neglected.

(*lower right*) Population topology: values of Mahalanobis' D computed from 3-sample pooled covariance matrix are shown within the enclosure. Outside are shown values computed from relevant 2-sample pooled covariance matrices and in brackets are values from which the growth components, as estimated by direction cosines for the largest principal component of sample covariance matrices, have been removed. Variates listed are those that contribute increases in $D^2$ that are significant at the 1% level.

Figure 12 shows that the collection from Escornebéou (F100718) is closer to that from the lower Aquitanian (F100706) than to that from the middle Aquitanian (F100700). On the hypothesis that the lineage exhibited irreversible trends in morphology and the information that F100706 is stratigraphically inferior to F100700, the location of F100718 in Fig. 12 indicates that, in turn, it is likely to be stratigraphically inferior to the lower Aquitanian horizon represented by F100706. Tests of the hypothesis that pairs of mean vectors are the same, indicate that the distance between Saucats collections is not significant at the 5% level: they appear to be samples drawn from a common population. But the distance between F100706 (Saucats, lower Aquitanian) and F100718 (Escornebéou) suggests that they are probably drawn from distinct populations ($0{\cdot}5 > P > 0{\cdot}01$); this is clearly established for F100700 (Saucats, middle Aquitanian): F100718 ($P < 0{\cdot}01$) Note that components of inter-sample distances due to differences in growth (Burnaby 1966) are small (Fig. 12).

A lineage is viewed here as a continuum of intergrading populations. The topology and tests of significance of differences between mean vectors indicate extent of morphological change between transects. Characters contributing to the distance between samples were detected by a procedure for computation of Mahalanobis' $D^2$ variate by variate (Rao 1952). Relevant 2-sample pooled coveriance matrices were used.

1. F100718 (Escornebéou) and F100706 (Saucats, lower Aquitanian). The principal contributors are width of the supplementary aperture ($X8$), horizontal translation of the penultimate whorl ($X11$) and basal diameter of the final chamber ($X13$). Mean values for these variates are higher in F100718 than in F100706.

2. F100718 (Escornebéou) and F100700 (Saucats, middle Aquitanian). Variates $X4$–$X9$, $X11$, $X13$ all provide increases in distance that are significant at the 1% level. Mean dimensions of apertures (except for $X8$) are larger in F100718 than in F100700 as are height of the spire and horizontal translation of the penultimate whorl. But the mean for the variate that contributes the largest increase in distance ($X13$, basal diameter of the final chamber) is larger in F100700 than in F100718.

3. F100706 (Saucats, lower Aquitanian) and F100700 (Saucats, middle Aquitanian). Principal contributors to inter-sample distance are heights of primary and supplementary apertures. The former decreases in mean value between F100706 and F100700, the latter increases.

## SYNTHESIS

1. All three collections are placed in the *Globigerinoides trilobus* lineage. Previous workers (Jenkins 1966; Blow 1969) have placed Saucats populations in this lineage. The hypothesis of a single lineage is supported by the extent to which specimens in the Escornebéou collection overlap the scatters for Saucats collections, and the absence of differences in patterns of intra-sample variation to which phylogenetic significance could be assigned. This

conclusion will require reconsideration as the ancestry of Neogene *Globigerinoides* is elucidated and criteria to detect parallelism and convergence emerge.

2. A mapping of samples indicates that the Escornebéou population represents a stage in the lineage prior to the Saucats populations. From this result the following trends in phenotypic variation are suggested (Fig. 1–10, 12):

(i) Reduction in size of apertures. This is well defined in the comparison between F100718 and F100700.

(ii) A more compact test. Saucats specimens tend to have a lower spire ($\times 9$), and horizontal translation of the penultimate whorl ($\times 11$) is reduced relative to Escornebéou material.

(iii) Increased overlap of the final chamber, and probably earlier chambers, on their predecessors. Chamber shape (spiral view) in F100718 and F100706 is akin to a truncated sphere. Increase in overlap extends the truncation. In F100700 chamber shape approaches hemispherical: the difference between basal ($\times 13$) and maximum ($\times 14$) diameters is reduced.

The mechanisms underlying these trends may only be suggested. The earliest population studied (Escornebéou) seems to have possessed greatest intrinsic variability. A possible model is one in which selection acted upon a wide field of phenotypic variation and moved it directionally while at the same time acting centripetally to decrease the intra-population scatter especially in the direction of growth. With the present data the field of variation of the highest sample still lies largely within the confines of the lowest but is more restricted.

3. Butt's conclusion that strata at Escornebéou are inferior to the Aquitanian stratotype is supported. Close coincidence between the first evolutionary appearance of the *Globigerinoides trilobus* stock (the *Globigerinoides* datum) and the base of the Aquitanian stratotype, recorded by Blow (1969), appears to be incorrect. In the Aquitaine basin there are earlier populations as at Escornebéou. Even the latter may not necessarily approximate the level of branching of the stock from an ancestral *Globigerina*.

4. As a criterion for correlation of the base of the Aquitanian Stage (as at the stratotype) the first occurrence of *Globigerinoides trilobus primordius* appears to be unreliable. However, within the Aquitaine basin the existence of trends in morphology may permit discrimination (with measurements analogous to those applied here) of earlier populations from those of the Aquitanian stratotype.


## ACKNOWLEDGMENTS

Dr P. N. Webb kindly made collections available for study and with Drs D. G. Jenkins and N. de B. Hornibrook commented on the manuscript.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# LETTERS TO THE EDITOR

## POST-OTIRAN MORAINES IN CANTERBURY

### COMMENT

A recent list of New Zealand radiocarbon age measurements (Grant-Taylor and Rafter 1971, p. 384) included a sample (S80/506) with a date ($NZ^{14}C548$) of 8,460 ± 120 yr B.P. collected by A. C. Beck in 1963 from the Macaulay valley above Lake Tekapo, Canterbury. The site description is given as "peat on fan debris and overlain by moraine".

I investigated the site (Fig. 1) in March 1967, and am doubtful that it records a glacial event. The carbonaceous material is, in fact, the A horizon of an undisturbed forest or subalpine scrub soil with wood and abundant cladodes of *Phyllocladus alpinus* present. It is buried by unsorted debris containing angular pebbles, cobbles and boulders, intermingled with finer material. I do not think this is glacial till for the following reasons. The setting is such that if ice had advanced over vegetation it might be expected, firstly, that the soil would be much disturbed, and secondly, that when the ice vacated the site a layer of ground moraine would be left. There is no sign of rock flour in the material burying the soil and plant remains, nor are the edges of the stony materials smoothed or worn.

Aerial photographs (3726/26, 27) are not very useful in distinguishing land forms in the locality. However, faint lines of ridges appear to radiate south-eastward from a point near the steep hillside to the north-west and these may indicate the direction of flow of a landslide. The tussock-covered surface above the dated deposit is hummocky and there are large angular blocks projecting from the soil, but these features could have originated in a landslide.

About 350 m upstream from the sample site lie the remnants of a large terminal moraine of the expanded Macaulay glacier. This moraine apparently was once more extensive but has been partly destroyed by the Macaulay River and Tom's Creek. It appeared to me, in the field, that the landslide overlay the moraine on its south-western margin, but there was no clear distinction between the two. The setting resembles very closely others in the Cameron and South Ashburton watersheds in Canterbury, quite near similar moraines. Wood and forest or subalpine scrub soils are buried by angular debris in these valleys, also, and I am confident that all of these deposits resulted from landslides or dirty snow avalanches. In the South Ashburton valley, wood buried by a landslide is dated (NZ1289) at 5,280 ± 105 yr B.P.

Despite this, the dating of the Macaulay site has value in providing a minimum age for vegetation development in the area and for establishing that the Macaulay glacier did not extend further down-valley later than about 8,400 years ago.

McGregor (1967) proposed the correlation of the Macaulay "moraine" with the Birch Hill moraine in the Tasman valley. I believe that we must reject this correlation for the reasons given above. Furthermore, because of local variations in the geomorphic history of different areas, moraines in different valley systems cannot be correlated with one another on morphological grounds, as was done by McGregor. I suggest that local chronologies

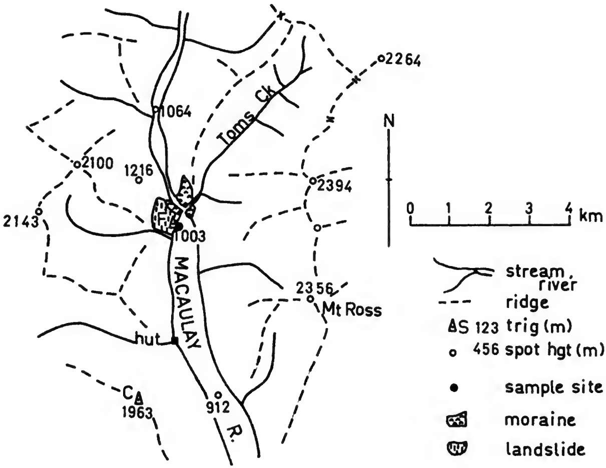


FIG. 1—The Macaulay valley and the sample site.

should be erected until such time as there is an adequate network of absolute dates. There are, at present, no dates which would allow us to be certain of the precise times of formation of the prominent post-Otiran moraines that are present in some Canterbury and Westland valleys. This applies also to the age of "approximately 10,000 years" suggested by Warren (1967) for the Waiho Loop near the Franz Josef glacier. Radiocarbon dates associated with such moraines are given below (Table 1).

In the Cameron valley, Arrowsmith Range (Table 1), 4·3 km from the existing glacier, a multiple set of moraines was formed more than 9,250 years ago. Its formation is likely to have occurred considerably less than 14,000 years ago, the date usually accepted as the end of the Otiran glaciation (Suggate and Moar 1970). It is also likely to have occurred at least several hundred years before 9,520 yr B.P. because a mature forest containing *Phyllocladus alpinus* occupied the upper Cameron valley then. It was overwhelmed, possibly originally as a result of seismic activity.

N.B. The present account was written after a field reappraisal of the Cameron valley site. Note that the site location and grid reference given in Grant-Taylor and Rafter (1971) are incorrect: the Cameron valley is in Canterbury, not Otago. The correct grid reference for NZ548 is S80/212506. There was an error in the field sheet from which A. C. Beck took the reference.

20 March 1972

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TABLE 1—Radiocarbon Dates Associated with Moraines

| DATING NUMBER ($NZ^{14}C$) | RADIOCARBON DATE (YR B.P.) | PLACE AND GRID REFERENCE | POSITION OF SAMPLE SITE WITH RESPECT TO MORAINE | SIGNIFICANCE | REFERENCE |
|---|---|---|---|---|---|
| NZ688 | 9,520 ± 95 | Cameron valley, Arrowsmith Range, Canterbury S73/597738 | The sample lies over till and under thick alluvial-colluvial deposits and the site lies some 150 m upstream of a prominent terminal moraine (Sample S73/549) | A major advance of the Cameron glacier occurred more than 9,520 years ago | Grant-Taylor and Rafter (1971, p. 383) |
| NZ548 | 8,460 ± 120 | Macaulay valley, above Lake Tekapo, Canterbury S80/212506 | As explained in the text above (Sample S80/506) | The Macaulay glacier has not extended further down-valley later than this date | Grant-Taylor and Rafter (1971, p. 384) Gair (1967) McGregor (1967) |
| NZ1289 | 5,280 ± 105 | Ashburton valley, Arrowsmith Range, Canterbury S73/575707 | Wood on alluvium, buried by landslide. The site lies some 2000 m downstream of a prominent terminal moraine (Sample S73/561) | The Ashburton glacier has not extended further down-valley later than this date | Unpublished. (Grant-Taylor pers. comm. 1972) |
| ——— | 5,120 | Birch Hill, Tasman valley, Canterbury ——— | Peat from a kettle-hole on the Birch Hill moraine surface | Provides a minimal date for the moraine | Gair (1967) (given as R. P. Goldthwait pers. comm.) |
| NZ724 | 2,430 ± 35 | Stony Creek, Waiho, Westland S71/835772 | Beneath outwash gravel near base of Waiho Loop moraine on its proximal side (Sample S71/508) | Provides a minimal date for the moraine | Grant-Taylor and Rafter (1971, p. 382) |

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## FURTHER COMMENT

The stratigraphic details of the site of sample S80/506 given by me contain an implication—that the terminal moraine in the Macaulay valley at Tom's Creek overlies the sampled wood—which should never have been made and was not in fact intended.

The deposit overlying the overwhelmed scrub is well described by Burrows. It is 3–4 ft thick where exposed towards its southern end and its surface is a relatively even surface sloping gently from west to east extending up to the downstream side of the moraine. Huge boulders protrude from the surface. I rejected the possibility that the deposit was a rock-filled snow avalanche as having no apparent reasonable source, and preferred an origin directly connected with the glacier. Burrows' mapping of the deposit as extending over the moraine makes the landslide origin compelling and I agree with him that the date ($NZ^{14}C548$) of 8,460 yr B.P. must now be regarded as a minimum age for this moraine.

However, it does not follow that McGregor's correlation of the Macaulay moraine with the Birch Hill moraines in the Tasman Valley must be rejected. Burrows rightly indicates there are dangers in using morphological and spatial relationships, but even so McGregor's correlation still seems a reasonable one. Especially so, considering that despite much investigation by Burrows and others there are still no direct datings of moraines.

To retain "local chronologies" until adequately dated by radiometric dates seems likely to proscribe correlation for ever.

19 April 1972

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# STRATIGRAPHIC POSITION OF THE NEW ZEALAND PLIOCENE-PLEISTOCENE BOUNDARY

## FURTHER COMMENT

In their reply to my comments concerning the New Zealand Pliocene-Pleistocene boundary (Jenkins 1971a) Vella and Nicol (1971) made a misleading statement, and further, they did not adequately answer my comments concerning either the status of (1) the Hautawan Stage in the Wairarapa or (2) the reliability of *Globorotalia truncatulinoides* as a marker for correlation.

### (1) *Hautawan Stage*

Vella and Nicol (1971) have singularly disagreed with my comments on the Hautawan because my statements were allegedly based solely on the work of Beu (1969). This is hardly true because in my discussion of the problem I quoted the work of Fleming (1953, 1959) and Boreham (1963) who had previously documented the diachronous nature of *Chlamys delicatula*.

Vella and Nicol (1970) had stated: "We shall continue to recognise the Hautawan in the Wairarapa as a climatic stage. . . ." and one of the published definitions of such a climate unit in geology states that it can have diachronous boundaries (American Commission on Stratigraphic Nomenclature, 1961). Thus a climate unit in geology is fundamentally different from a time-stratigraphic unit, but my suggestion that the Wairarapa Hautawan was similar to the climate unit (Jenkins 1971a) was rejected by Vella and Nicol (1971). In their reply it would have been opportune for them to clarify their meaning of a "climatic stage" because "stage" had been restricted in a number of stratigraphic codes to be the basic time-stratigraphic unit, and therefore a "climatic stage" is contrary to the recommended usage of "stage" in New Zealand (Hedberg 1961).

### (2) *Globorotalia truncatulinoides*

In the type Calabrian section, *G. truncatulinoides* has only been recorded from the *top* of the measured section (Bayliss 1969) and there is no published record of the evolutionary transition *Globorotalia tosaensis*–*G. truncatulinoides* in the section (Jenkins 1971b). Yet, its first appearance in the Mangaopari section is correlated with the base of the type Calabrian (Kennett *et al.* 1971) and therefore this correlation is still un-acceptable.

The main reason for regarding the initial appearance of *G. truncatulinoides* as an accurate marker for long distance correlation was based on the records of its evolutionary transition from *G. tosaensis* in deep-sea cores (Berggren *et al.* 1967; Glass *et al.* 1967). The evolutionary appearance coincided with the Olduvai event and not the beginning of the Gilsa as stated by Vella and Nicol (1971) and this does not support their conclusion regarding their attempted correlation of *G. truncatulinoides* that it ". . . greatly enhances its significance".

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I regarded it as very significant that Devereux *et al.* (1970) had recorded the range of *G. tosaensis* immediately preceding the *G. truncatulinoides* range in the Mangaopari section as though the evolutionary transition was present. I note that more recently the lower part of the *G. truncatulinoides* range has been referred "to the *G. crassaformis* complex" by Kennett *et al.* (1971). Thus the initial appearance of *G. truncatulinoides*, which ranges through to Recent, appears now to be cryptogenic and therefore has much less value for long distance correlation. The reliability of *G. truncatulinoides* for correlation even within New Zealand is questionable because of the recently published evidence of its occurrence in the North Island Pliocene (Jenkins 1971c).

Kennett *et al.* (1971) have published a valuable magnetic chronology for the Mangaopari section of Pliocene–Lower Pleistocene and fortunately the attitude of Vella and Nicol (1971) to the stratigraphic and paleogeographic distributions of *C. delicatula* and *G. truncatulinoides* will not diminish its value.

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